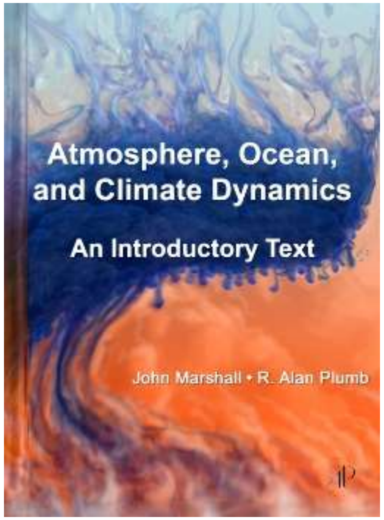
GEF 1100 – Klimasystemet

Chapter 7: Balanced flow



Prof. Dr. Kirstin Krüger (MetOs, UiO)

UiO : University of Oslo

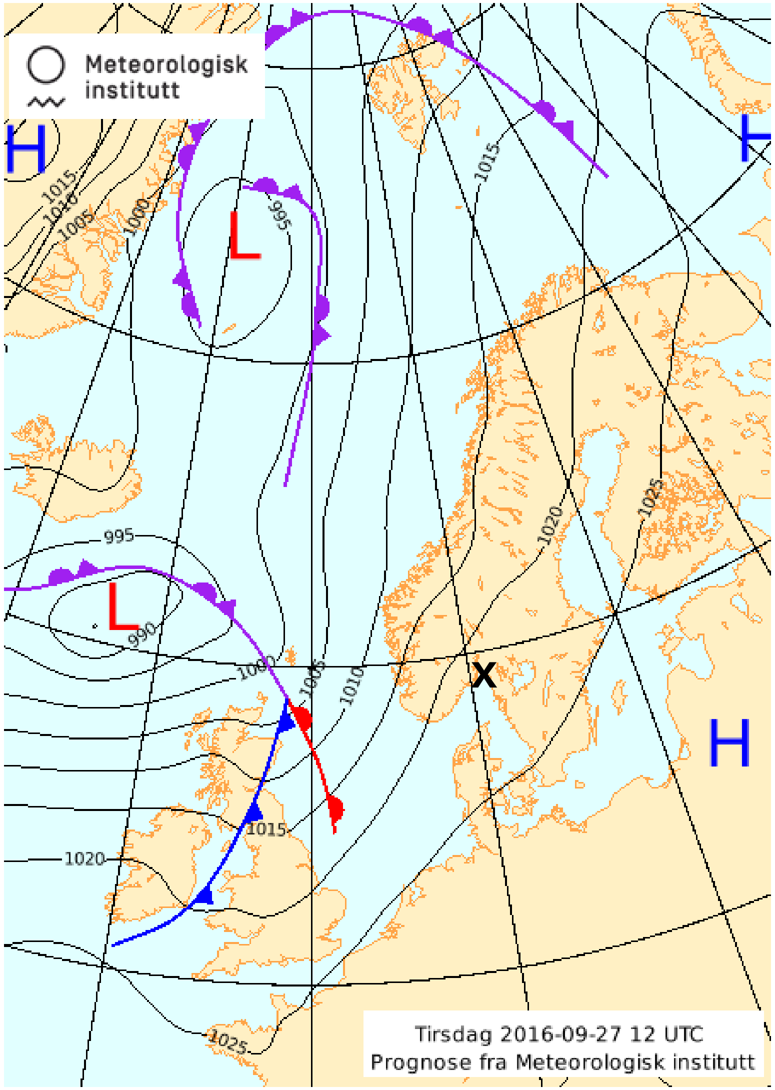
Lecture Outline – Ch. 7

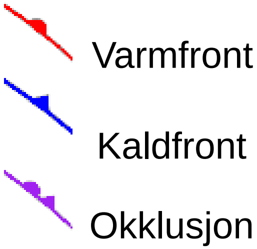
Ch. 7 – Balanced flow

1. Motivation
2. Geostrophic motion
    - 2.1 Geostrophic wind
    - 2.2 Synoptic charts
    - 2.3 Balanced flows
3. Thermal wind equation*
4. Subgeostrophic flow: The Ekman layer
    - 5.1 The Ekman layer*
    - 5.2 Surface (friction) wind
    - 5.3 Ageostrophic flow
5. Summary
6. Take home message

**With add ons.*

Today's weather chart

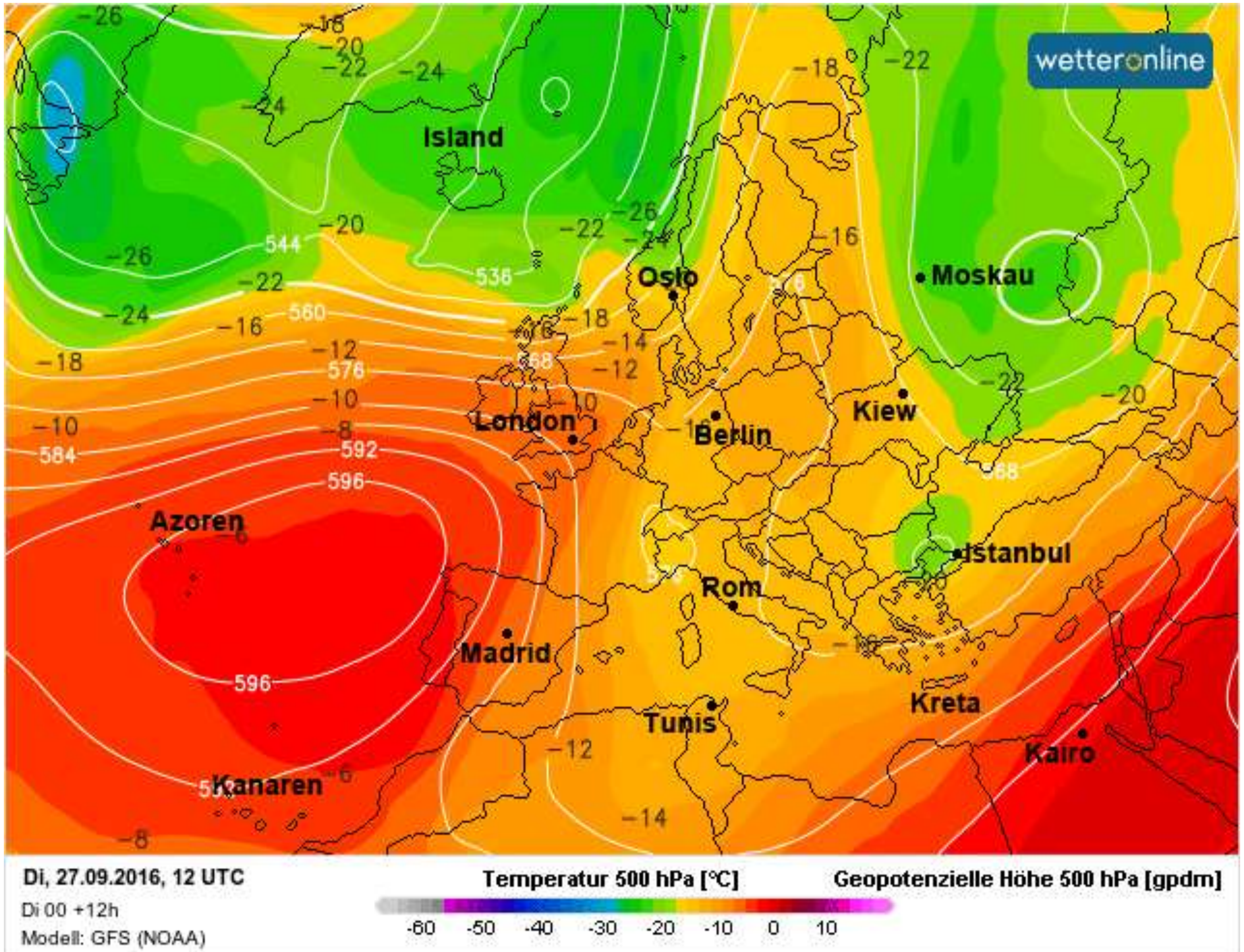




Where does the wind come from in Oslo?

www.yr.no

Today's 500 hPa (~5 km) weather map



www.wetteronline.de

GFS: Global Forecast System-Model

How is the wind blowing? Which forces are acting?

Scale analysis – geostrophic balance -1-

First consider magnitudes of the first 2 terms in momentum eq. for a fluid on a rotating sphere (Eq. 6-43 $\frac{D\mathbf{u}}{Dt} + f\,\hat{\mathbf{z}} \times \boldsymbol{u} + \frac{1}{\rho}\nabla p + \nabla\Phi = \mathcal{F}$) for horizontal components in a free atmosphere ($\mathcal{F}$=0,$\nabla\Phi=0$):

- $$\frac{D\mathbf{u}}{Dt} + f\hat{\mathbf{z}} \times \boldsymbol{u} = \underbrace{\frac{\partial u}{\partial t}}_{\frac{U}{T}\;\;10^{-4}} + \underbrace{\boldsymbol{u}\cdot\nabla\boldsymbol{u}}_{\frac{U^2}{L}\;\;10^{-4}} + \underbrace{f\hat{\mathbf{z}} \times \boldsymbol{u}}_{fU\;\;10^{-3}}$$

Large-scale flow magnitudes in atmosphere: u and $v \sim \mathcal{U}$, $\mathcal{U}$~10 m/s, length scale: L ~10^6 m, time scale: $\mathcal{T}$ ~10^5 s, $\mathcal{U}$ / $\mathcal{T}\approx \mathcal{U}^2$/L ~$10^{-4}$ ms^{-2}, $f_{45°}$ ~ 10^{-4} s^{-1}

- **Rossby number** R_0: - Ratio of acceleration terms (U^2/L) to Coriolis term (f U),
  - R_0=U/f L
  - $R_0\simeq$0.1 for large-scale flows in atmosphere ($R_0\simeq10^{-3}$ in ocean, see Chapter 9)

Scale analysis – geostrophic balance -2-

- Coriolis term is left (≙ smallness of R_0) together with the pressure gradient term (~10^{-3}):

$$f\hat{\mathbf{z}} \times \boldsymbol{u} + \frac{1}{\rho}\nabla p = 0$$ Eq. 7-2 ***geostrophic balance***

- It can be rearranged to ($\hat{\mathbf{z}} \times \hat{\mathbf{z}} \times \boldsymbol{u} = -\boldsymbol{u}$):

$$\boldsymbol{u}_g = \frac{1}{f\rho}\hat{\mathbf{z}} \times \nabla p$$ Eq. 7-3 ***geostrophic flow***

$$\hat{\mathbf{z}} \times \nabla p = \begin{vmatrix} \hat{x} & \hat{y} & \hat{z} \\ 0 & 0 & 1 \\ \frac{\partial p}{\partial x} & \frac{\partial p}{\partial y} & \frac{\partial p}{\partial z} \end{vmatrix} = \hat{x}\left(-\frac{\partial p}{\partial y}\right) + \hat{y}\left(\frac{\partial p}{\partial x}\right)$$

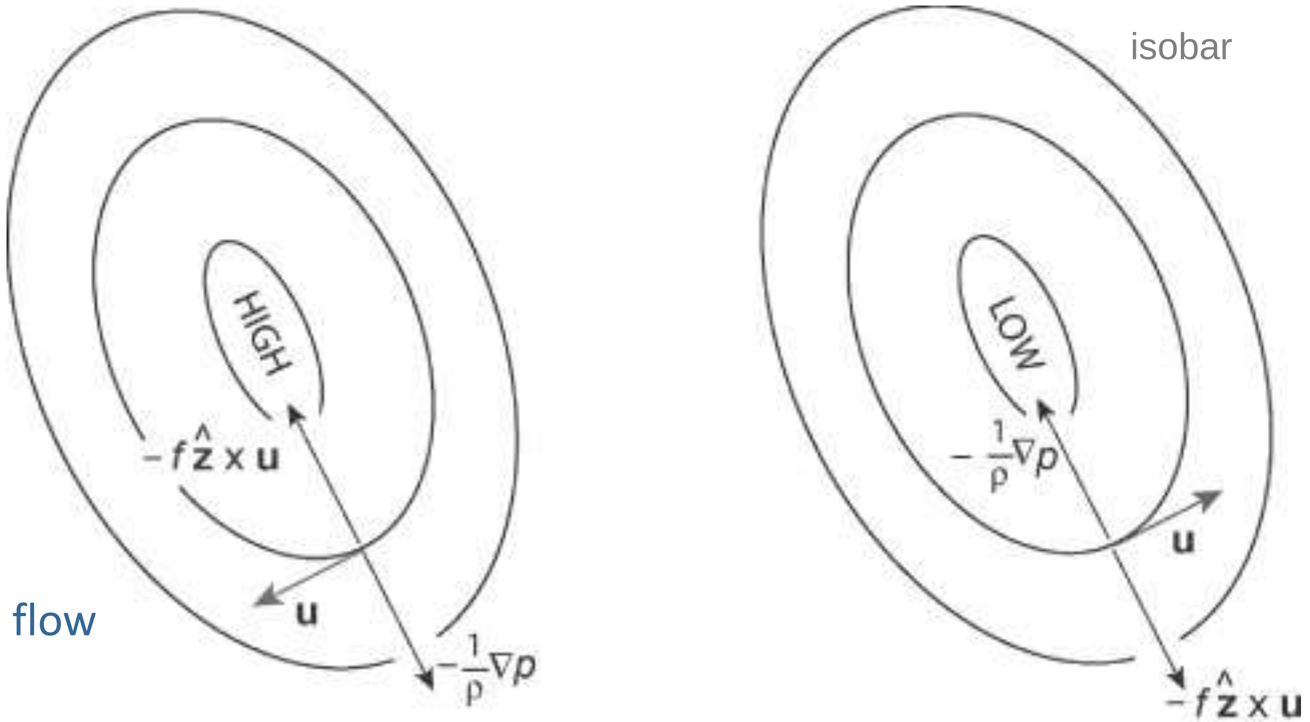
- In vector component form:

$$(u_g, v_g) = \left(-\frac{1}{f\rho}\frac{\partial p}{\partial y}, \frac{1}{f\rho}\frac{\partial p}{\partial x}\right)$$ Eq. 7-4

Note: For *geostrophic flow* the pressure gradient is balanced by the Coriolis term; to be approximately satisfied for flows of small R_0.

Geostrophic flow (NH: *f*>0)

High:
clockwise flow



Low:
anticlockwise flow

Figure 7.1: Geostrophic flow around a high pressure center (left) and a low pressure center (right). (Northern hemisphere case, $f > 0$.) The effect of Coriolis deflecting flow "to the right" (see Fig. 6.10) is balanced by the horizontal component of the pressure gradient force, $-1/\rho\, \nabla p$, directed from high to low pressure.

Marshall and Plumb (2008)

Geostrophic flow

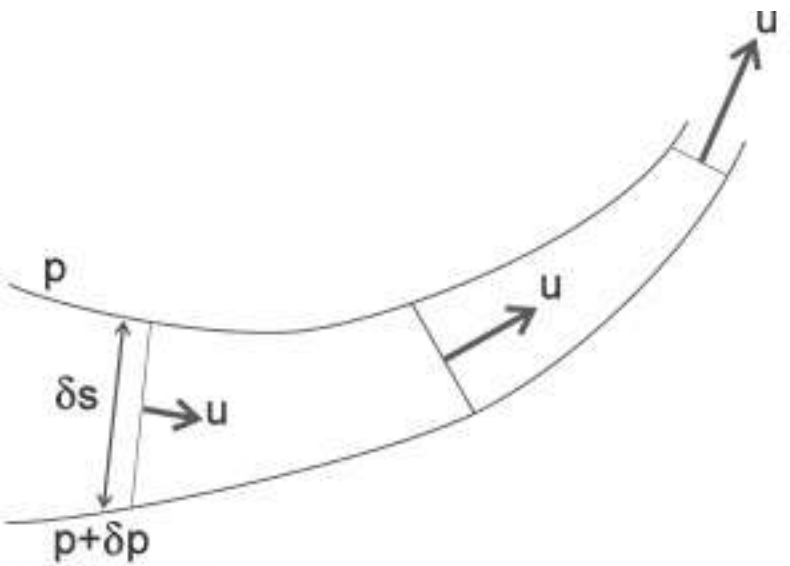


Figure 7.2: Schematic of two pressure contours (isobars) on a horizontal surface. The geostrophic flow, defined by Eq. 7-3, is directed along the isobars; its magnitude increases as the isobars become closer together.

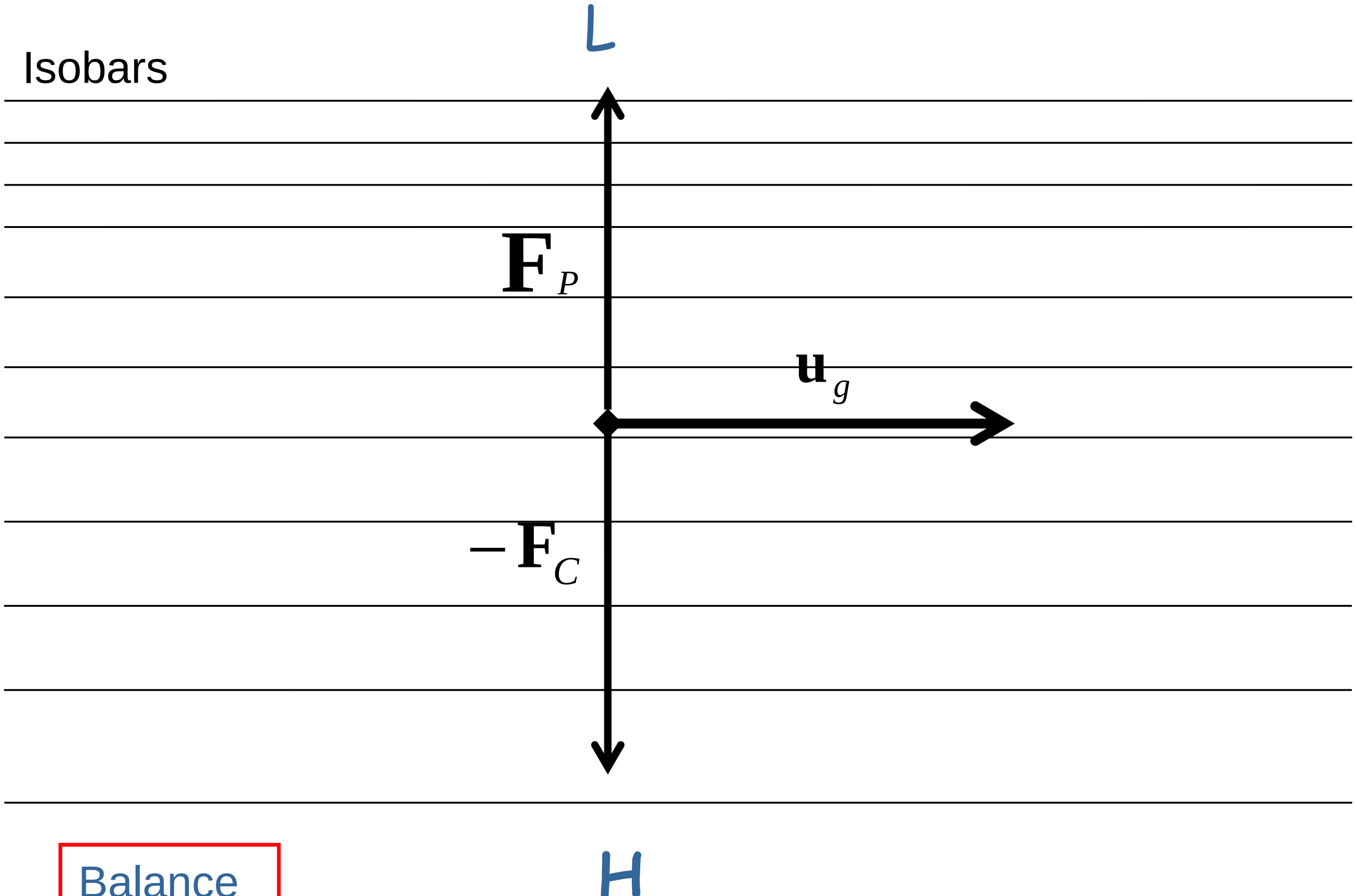
$$|\boldsymbol{u}_g| = \frac{1}{f\rho}|\nabla p| = \frac{1}{f\rho}\frac{\delta p}{\delta s}$$

s: distance

Marshall and Plumb (2008)

Note: The geostrophic wind flows parallel to the isobars and is strongest where the isobars are closest (pressure maps > winds).

Geostrophic wind

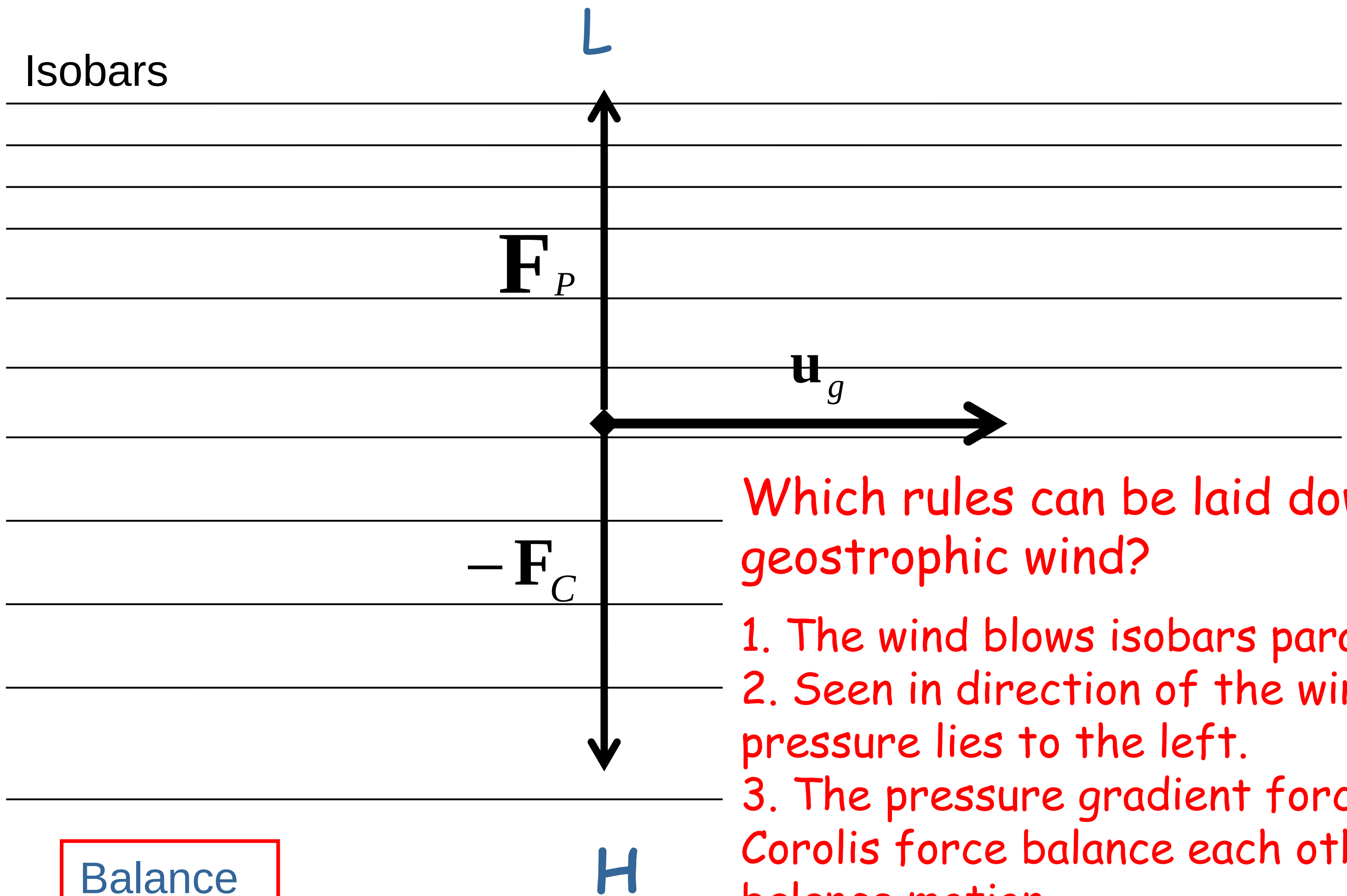


Balance between:

$\vec{F}_P = -\vec{F}_C$

$$\boldsymbol{u}_g = \frac{1}{f\rho}\hat{\boldsymbol{z}} \times \nabla p$$

Geostrophic wind



Which rules can be laid down for the geostrophic wind?

1. The wind blows isobars parallel.
2. Seen in direction of the wind, the low pressure lies to the left.
3. The pressure gradient force and the Corolis force balance each other ⇒ balance motion.

⇒ The geostrophic wind closely describes the horizontal wind above the boundary layer!

Balance between:

$\vec{F}_P = -\vec{F}_C$

Geostrophic flow – vertical component

Vertical component of geostrophic flow, as defined by Eq. 7-3, is zero. This can not be deduced directly from geostrophic balance (Eq. 7-2).

- Consider incompressible fluids (like the ocean):
  ρ can be neglected, $f \sim \text{const.}$ for planetary scales, then geostrophic wind components (Eq. 7-4) gives:

  $\frac{\partial u_g}{\partial x} + \frac{\partial v_g}{\partial y} = 0$ Eq. 7-5 ``Horizontal flow is non divergent.''

  Comparison with continuity equation $(\nabla \cdot \boldsymbol{u} = 0)$ => $w_g = 0$ => $\frac{\partial w_g}{\partial z} = 0$

  => geostrophic flow is horizontal.

- For compressible fluids (like the atmosphere) ρ varies, see next slide.

Geostrophic wind – pressure coordinates "$_p$"

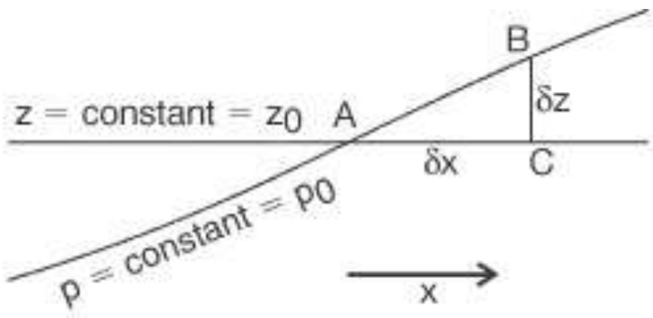


Figure 7.3: Schematic used in converting from pressure gradients on height surfaces to height gradients on pressure surfaces.

Marshall and Plumb (2008)

We derive from Eq. 7.3 in pressure coordinates:

$$\boldsymbol{u}_g = \frac{g}{f}\hat{\boldsymbol{z}}_p \times \nabla_p z \qquad \text{Eq. 7-7}$$

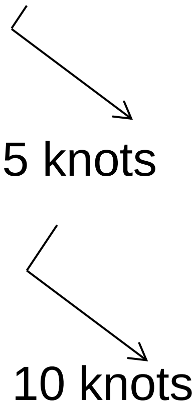
$$(u_g, v_g) = \left(-\frac{g}{f}\frac{\partial z}{\partial y}, \frac{g}{f}\frac{\partial z}{\partial x}\right) \qquad \text{Eq. 7-8}$$

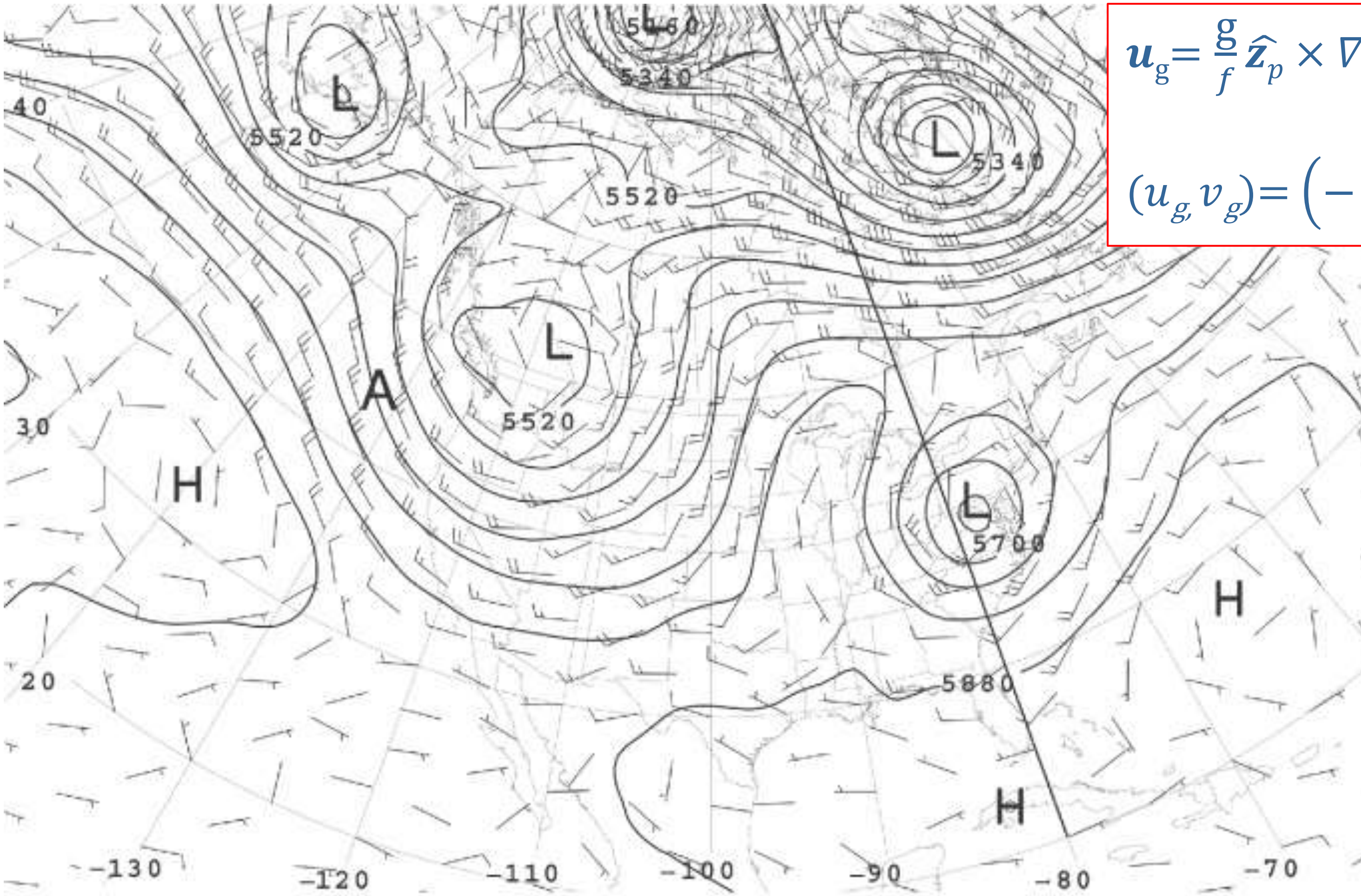
$$\nabla_p \cdot \boldsymbol{u}_g = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} = 0 \qquad \text{Eq. 7-9}$$

Geostrophic wind is non-divergent in pressure coordinates, if $f \sim \text{const.}$ ($\leq$ 1000 km).

Note: Geopotential height (*z*) contours are streamlines of the geostrophic flow on pressure surfaces; geostrophic flow streams along *z* contours (as *p* contours on *height* surfaces) (Fig. 7-4).

500 hPa wind and geopotential height (gpm), 12 GMT June 21, 2003 over USA





$$\boldsymbol{u}_g = \frac{g}{f}\hat{\boldsymbol{z}}_p \times \nabla_p z$$

$$(u_g, v_g) = \left(-\frac{g}{f}\frac{\partial z}{\partial y}, \frac{g}{f}\frac{\partial z}{\partial x}\right)$$

Note: Geopotential height ($z_{GH} \approx z$*; see Chapter 5 Eq. 5.5*) contours are streamlines of the geostrophic flow on pressure surfaces; geostrophic flow streams along *z* contours (as *p* contours on *height* surfaces) (Fig. 7-4).

Marshall and Plumb (2008)

Summary: The geostrophic wind $\boldsymbol{u}_g$

- Geostrophic wind: “wind on the turning earth“: Geo (Greek) “Earth“, strophe (Greek) “turning”.
- The friction force and centrifugal forces are negligible.
- Balance between pressure gradient and Coriolis force
  → **geostrophic balance**
- The geostrophic wind is a horizontal wind.
- It displays a good approximation of the horizontal wind in the free atmosphere; not defined at the equator:
  - Assumption: $u=u_g$ $v=v_g$

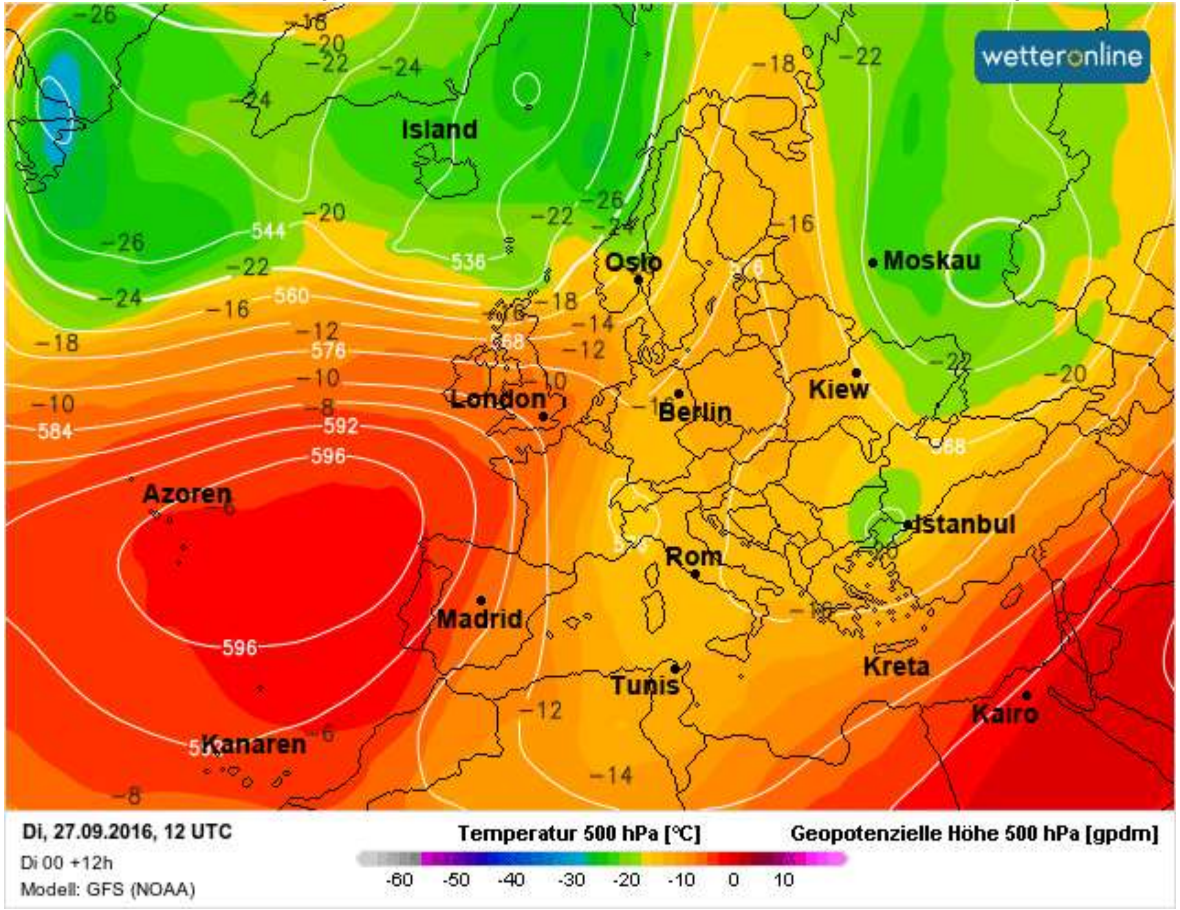
$$\Rightarrow \boldsymbol{u_g} = \frac{1}{f\rho}\hat{\mathbf{z}} \times \nabla p \qquad u_g = -\frac{1}{f\rho}\frac{\partial p}{\partial y} \quad v_g = \frac{1}{f\rho}\frac{\partial p}{\partial x}$$

ρ: density
$f = 2\,\Omega \sin\phi$;
at equator $f = 0$
at Pole f is maximum

Quiz

How is the wind blowing? Where are the low and high pressure systems? Which forces are acting?

Today‘s 500 hPa (~5 km) weather map



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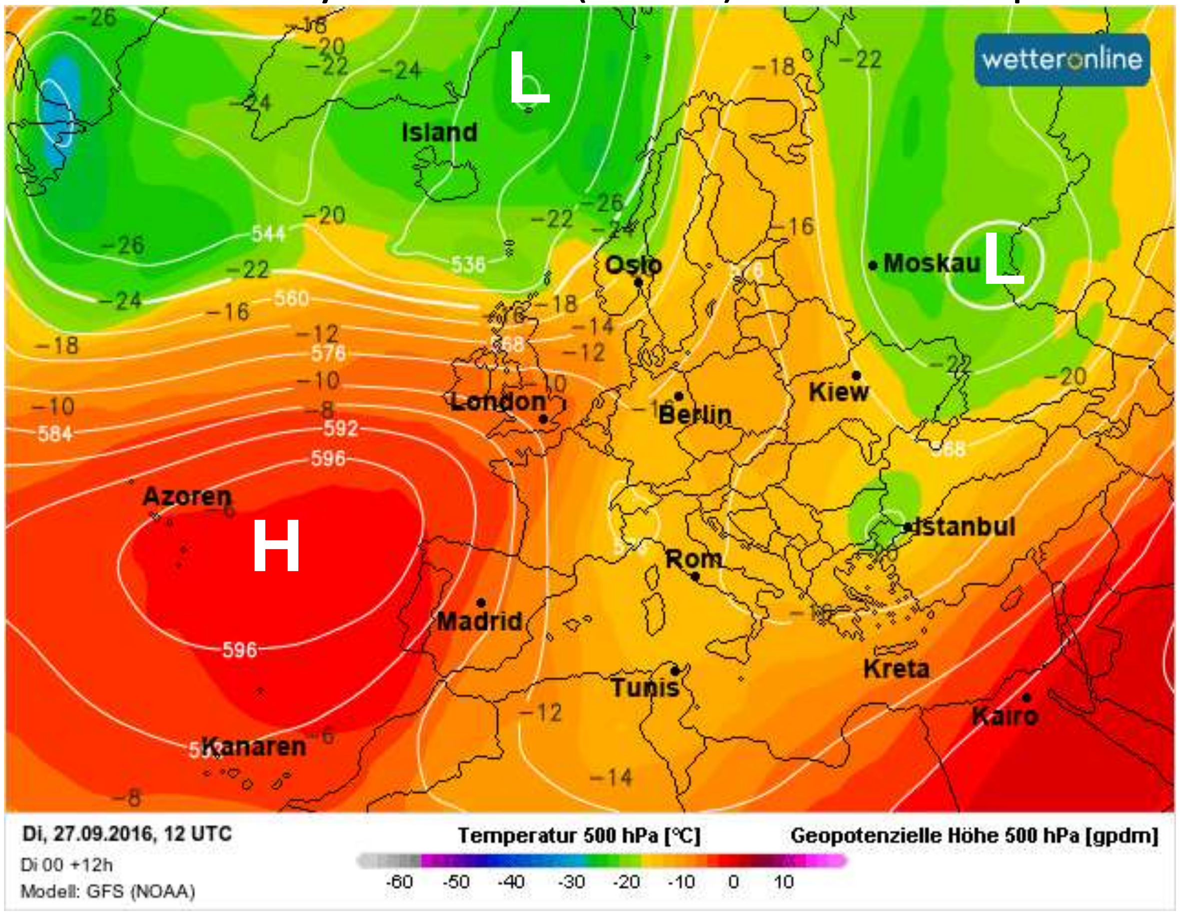
GFS: Global Forecast System-Model

How is the wind blowing? Which forces are acting?

Quiz

How is the wind blowing? Where are the low and high pressure systems? Which forces are acting?

Today's 500 hPa (~5 km) weather map



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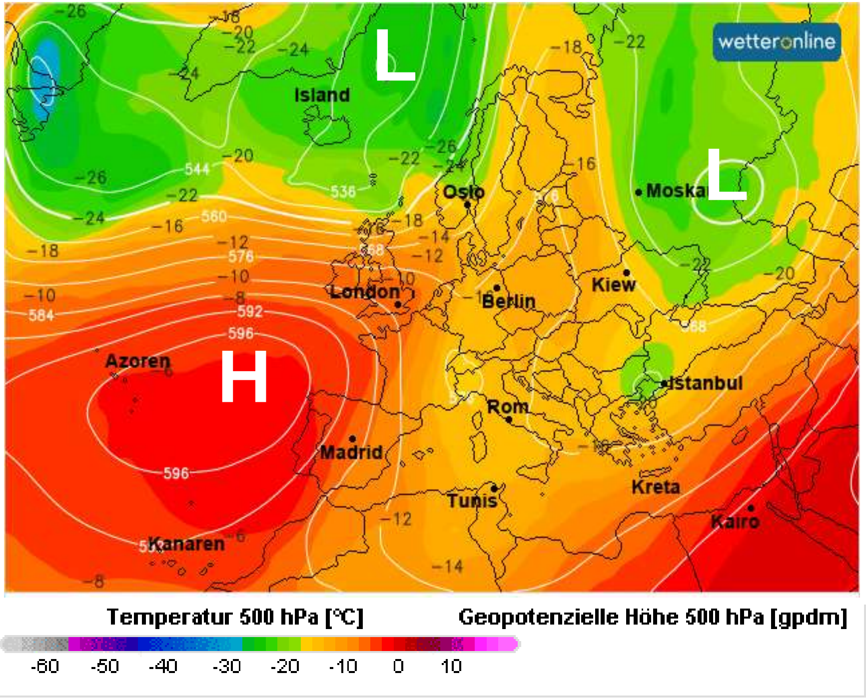
GFS: Global Forecast System-Model

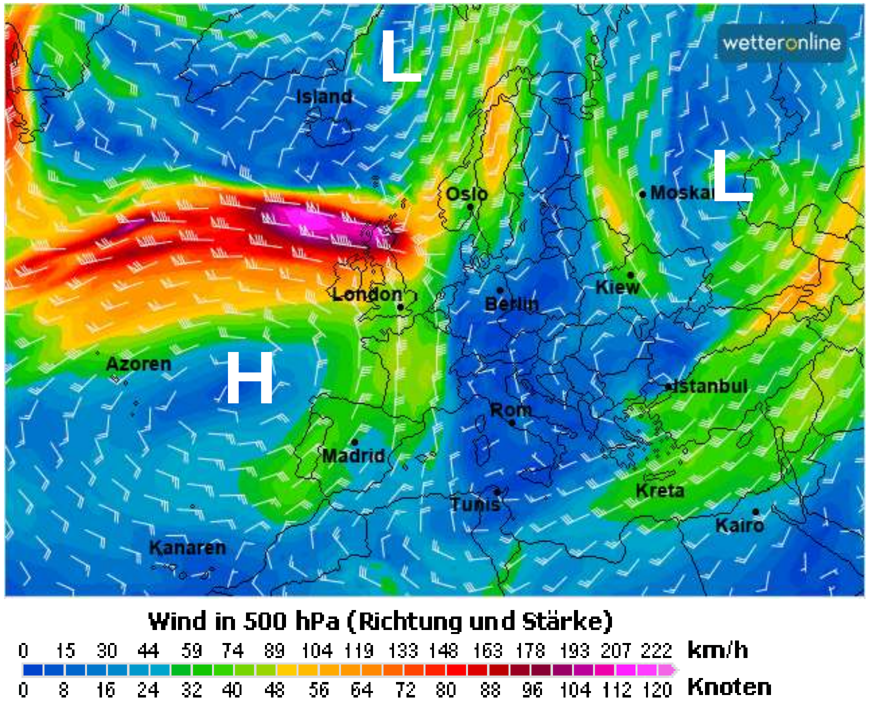
How is the wind blowing? Which forces are acting?

[Quiz](#)

How is the wind blowing? Where are the low and high pressure systems? Which forces are acting?

Tues 27.09.2016 12 UTC, GFS model





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The geostrophic wind $\mathbf{u}_g$

- **Geostrophic wind balance (GWB)**:

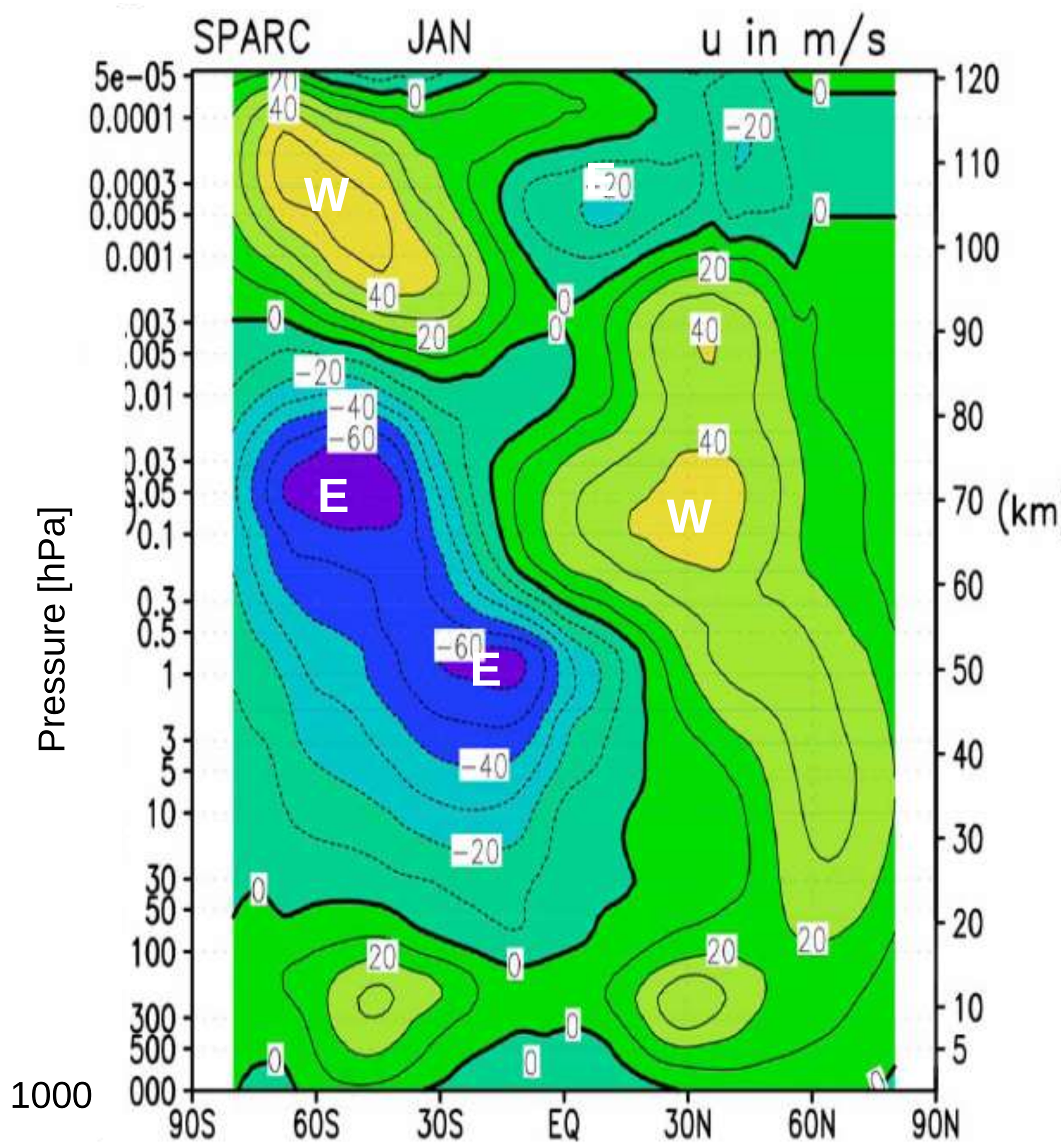
  Simultaneous wind and pressure measurements in the open atmosphere show that mostly the GWB is fully achieved.

  Advantage: The wind field can be determined directly from the pressure or height fields.

  Disadvantage: The horizontal equation of motion becomes purely diagnostic; stationary state.

Zonal mean
zonal wind u (m/s)

January average
SPARC climatology



Randel et al 2004

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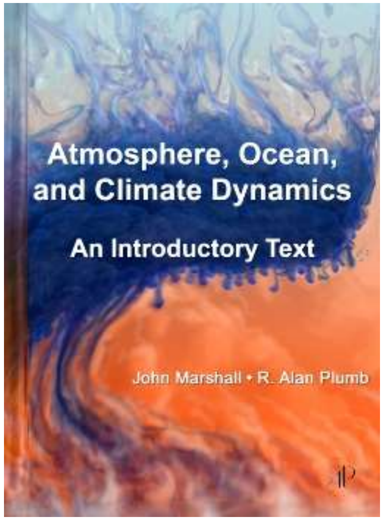
Chapter 7: Balanced flow



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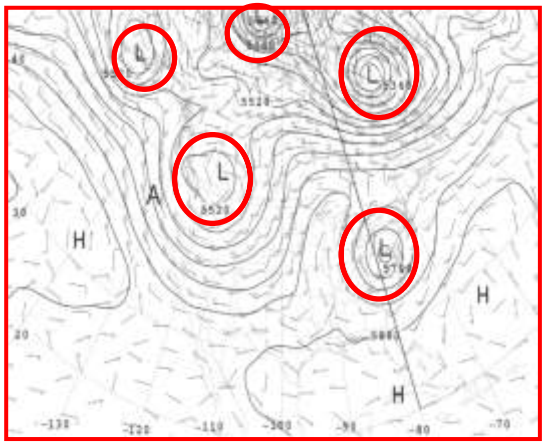
**With add ons.*

Rossby number R_0 @500hPa, 12 GMT June 21, 2003



$R_0 > 0.1$

- R_0 : Ratio of acceleration term ($U/T \approx U^2/L$) to Coriolis term (f U)
- $R_0 = U/f L$
- $R_0 \simeq 0.1$ for large-scale flows in atmosphere

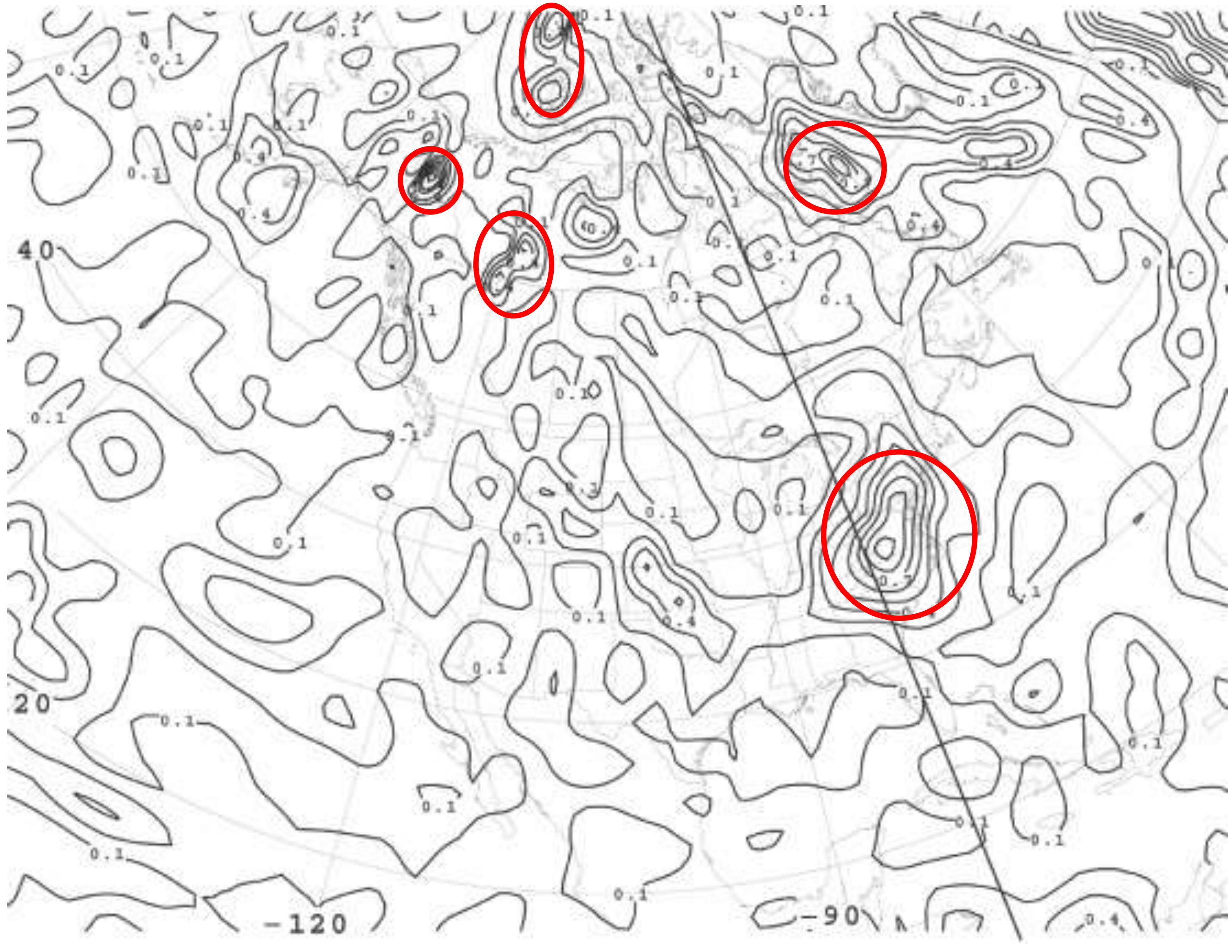


Figure 7.5: The Rossby number for the 500-mbar flow at 12 GMT on June 21, 2003, the same time as Fig. 7.4. The contour interval is 0.1. Note that Ro ~ 0.1 over most of the region but can approach 1 in strong cyclones, such as the low centered over 80° W, 40° N.

Marshall and Plumb (2008)

(Other) Balanced flows

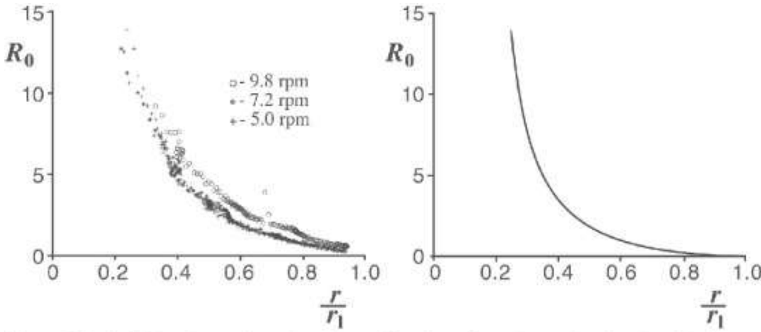


Figure 7.6: Left: The R_o number plotted as a function of nondimensional radius (r/r_1) computed by tracking particles in three radial inflow experiments (each at a different rotation rate, quoted here in revolutions per minute [rpm]). Right: Theoretical prediction based on Eq. 7-12.

Marshall and Plumb (2008)

- **Geostrophic balance** holds, when small positive R_0 (~ 0.1).

Pressure gradient force and Coriolis force are in balance.

- **Gradient wind balance** (R_0~1)

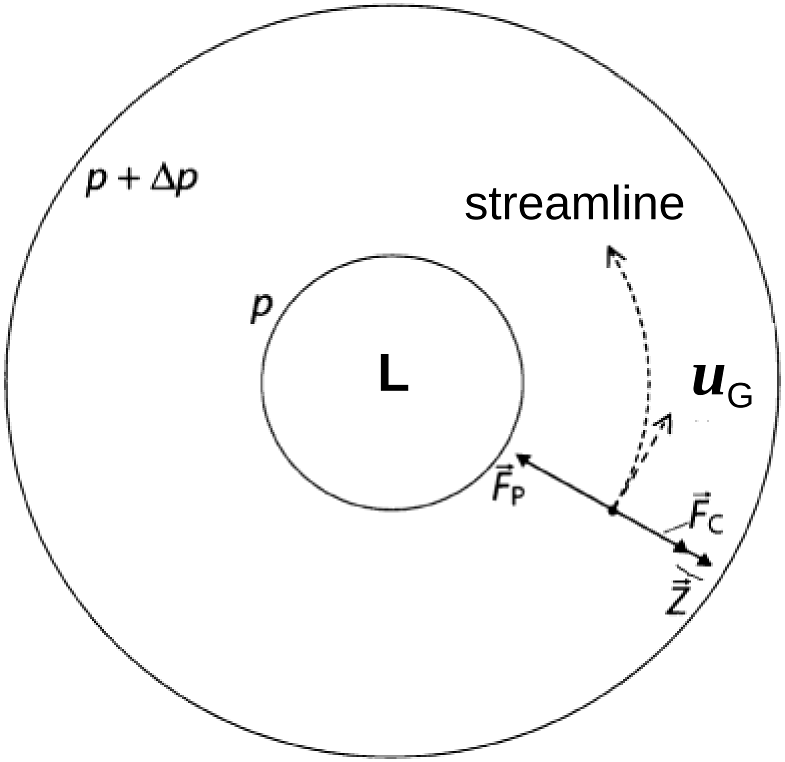
Pressure gradient force, Coriolis force and Centrifugal force play a role.

- **Cyclostrophic wind balance** (R_0>1)

Pressure gradient force and Centrifugal force play a role.

Gradient wind $\boldsymbol{u}_G$ – 2 cases with same $\boldsymbol{F}_p$

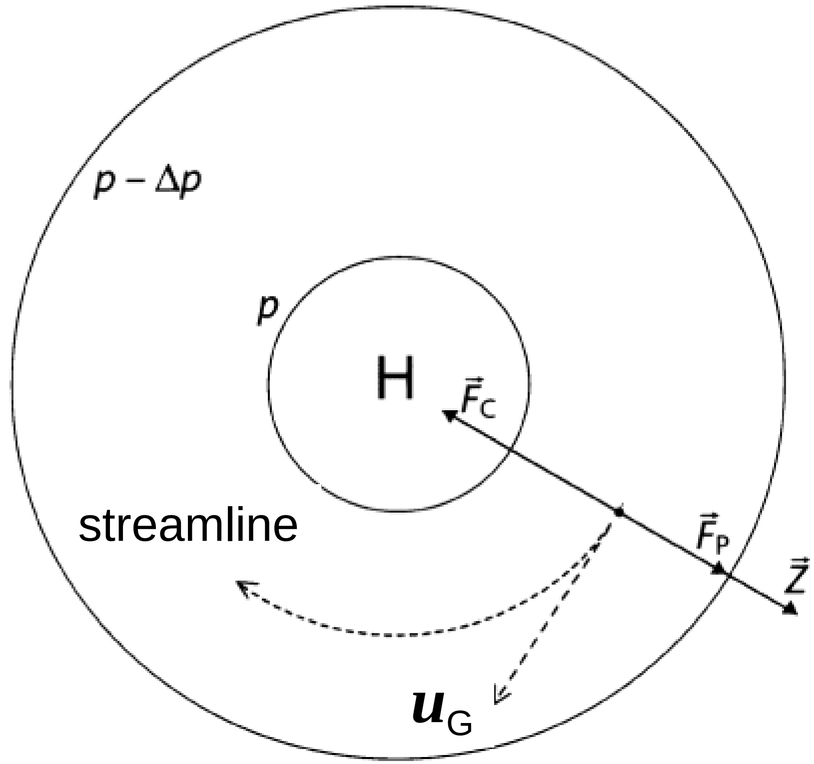
Case 1: sub-geostrophic



Balance between the forces:

$\vec{F}_{pressure} = -(\vec{F}_{Coriolis} + \vec{F}_{Z(centrifugal)})$

Case 2: super-geostrophic

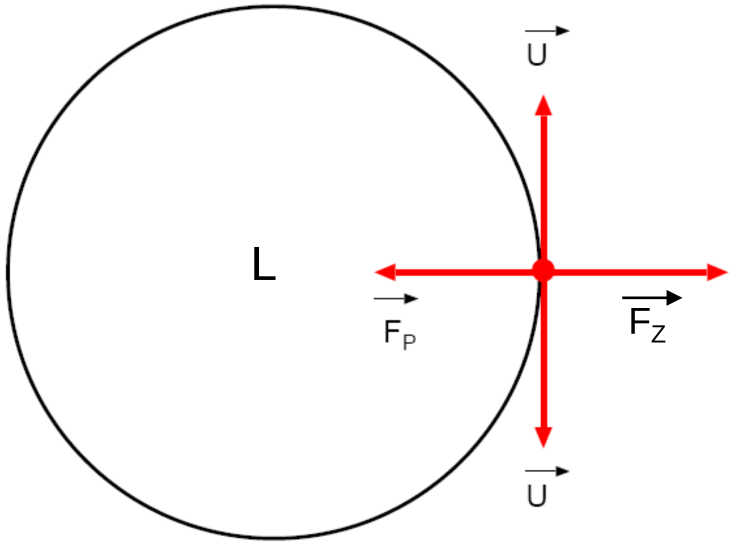


Balance between the forces:

$\vec{\mathbf{F}}_{pressure} + \vec{\mathbf{F}}_{Z(centrifugal)} = -\vec{\mathbf{F}}_{Coriolis}$

Kraus (2004)

Cyclostrophic wind



Balance between:

$\mathbf{F}_P = -\mathbf{F}_Z$

- Physically appropriate solutions are orbits around a low pressure centre, that can run both cyclonically (counter clockwise) as well as anticyclonically (clockwise).
- The cyclostrophic wind appears in **small-scale whirlwinds** (dust devils, tornados).

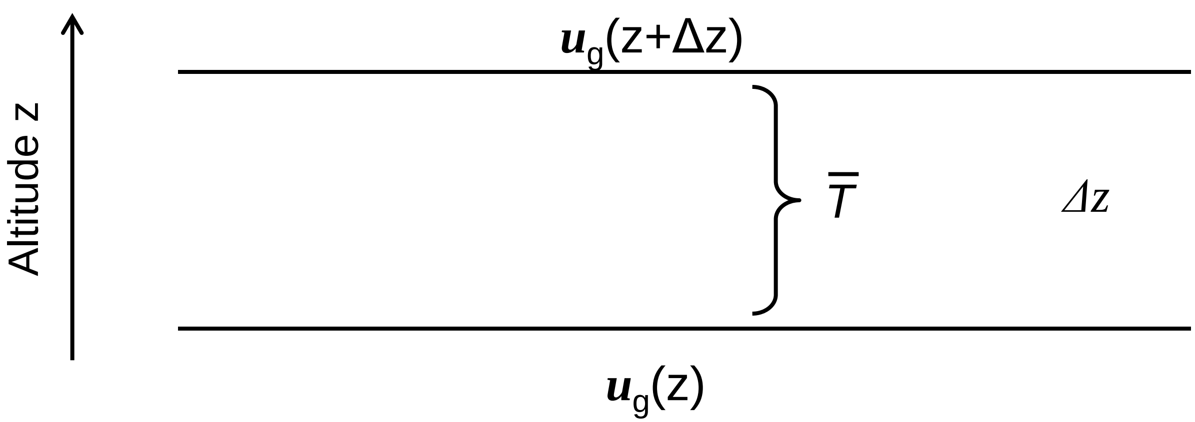
The thermal wind

Geostrophic wind change with altitude z?

→"Thermal wind" ($\Delta_z u_g$)

Applying geostrophic wind equation together with ideal gas law (Eq. 1-1 *p=ρRT*); insert finite vertical layer Δz* with average temperature $\overline{T}$, leads to:
*(*Δ: finite difference)*

$$\Delta_z \boldsymbol{u}_g = \frac{1}{\overline{T}}(\Delta_z T)\boldsymbol{u}_g + \frac{g}{f\overline{T}}\Delta z\,\hat{\mathbf{z}} \times \nabla_h \overline{T}$$



$\boldsymbol{u}_g$: geostrophic wind vector, *T:* temperature *ρ*: density *p:* pressure; *R*: gas constant for dry air,
g: gravity acceleration, *f* : *Coriolis parameter, z: geometric height,* $\hat{\mathbf{z}}$: unit vector in z−direction

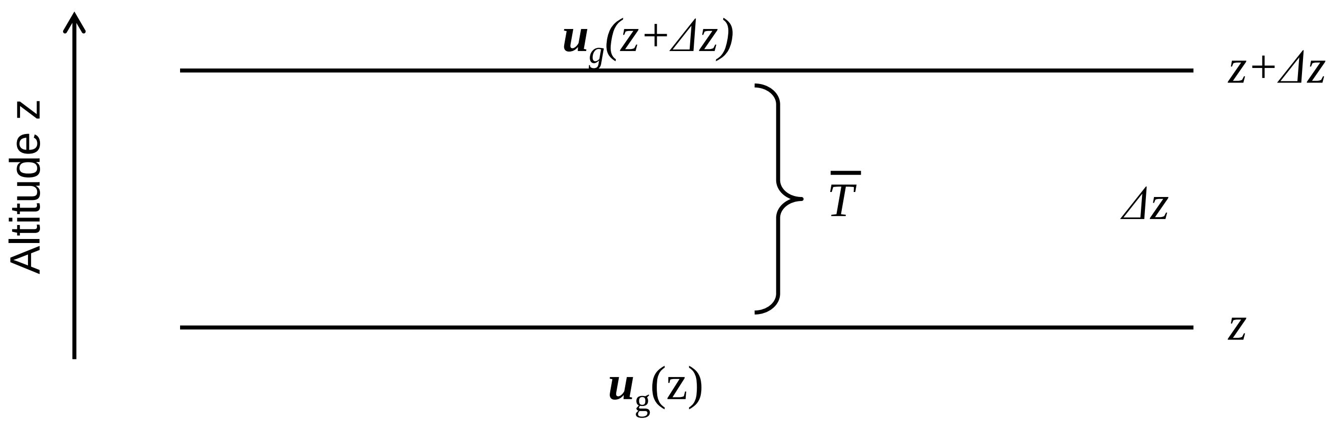
Thermal wind approximation

$$\Delta_z \boldsymbol{u}_g = \cancel{\frac{1}{\overline{T}}(\Delta_z T)\boldsymbol{u}_g} + \frac{g}{f\overline{T}}\Delta z\,\hat{\mathbf{z}} \times \nabla_h \overline{T}$$

In many cases: 1) $\boldsymbol{u}_g$<10 m/s; or 2) strong horizontal temperature gradient; then the first term can be neglected.

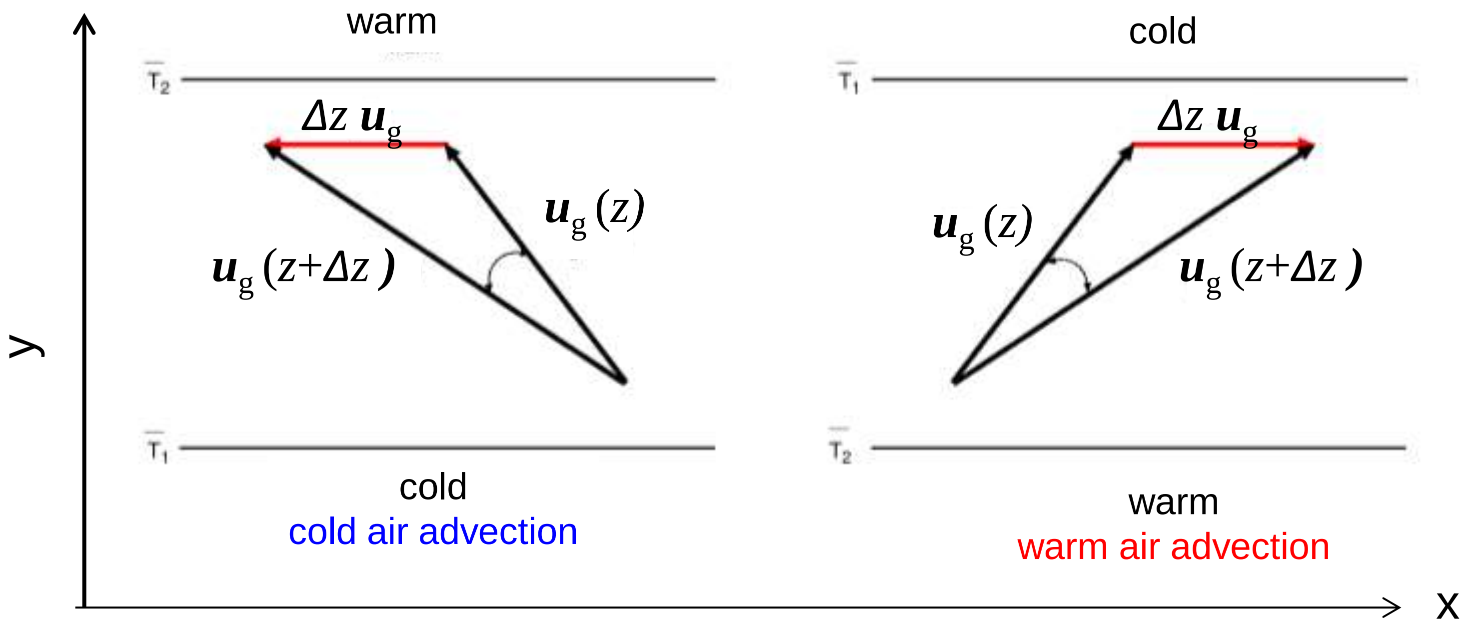
$$\Delta_z \mathbf{u}_g = \mathbf{u}_g(z+\Delta z) - \mathbf{u}_g(z) \approx \frac{g}{f\overline{T}}\Delta z\ \hat{\mathbf{z}} \times \nabla_h \overline{T}$$

$$\Rightarrow \Delta_z \mathbf{u}_g \approx \hat{\mathbf{z}} \times \nabla_h \overline{T}$$



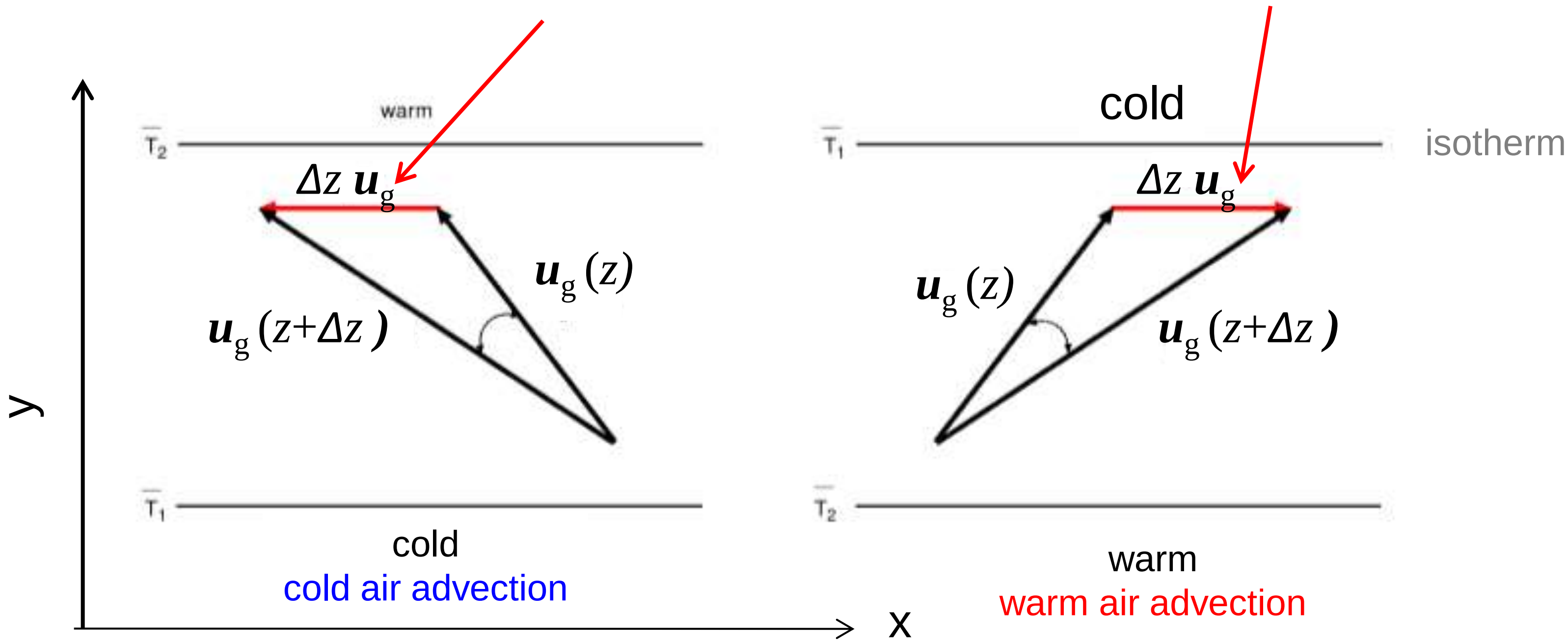
Note: In this approximation, the wind change is always parallel to the average isotherms (lines of const. temperature).

Thermal wind approximation



Which rules can I deduce from the thermal wind approximation?

Thermal wind approximation



Which rules can I deduce from the thermal wind approximation, if $u_g<10\ m/s$?

- The geostr. wind change with height **is parallel to the mean isotherms**,
- whereby the **colder air remains to the left on the NH**,
- **geostrophic left rotation with increasing height for cold air advection**,
- **geostrophic right rotation with increasing height for warm air advection**.

Thermal wind:

$$\frac{\partial \boldsymbol{u}_g}{\partial z} = \frac{\alpha g}{f} \hat{\boldsymbol{z}} \times \nabla T \quad \text{Eq. 7-18}$$

$$\begin{cases} \frac{\partial u_g}{\partial z} = u_g(z+\Delta z) - u_g(z) \approx -\frac{\alpha g}{f}\frac{\partial T}{\partial y} \\ \frac{\partial v_g}{\partial z} = v_g(z+\Delta z) - v_g(z) \approx \frac{\alpha g}{f}\frac{\partial T}{\partial x} \end{cases}$$

α: Thermal expansion coefficient (Ch. 4)
g: Gravity acceleration

Geostrophic wind:

$$\boldsymbol{u}_g = \frac{1}{f\rho} \hat{\boldsymbol{z}} \times \nabla p \quad \text{Eq. 7-3}$$

$$\boldsymbol{u}_g = \frac{g}{f} \hat{\boldsymbol{z}}_p \times \nabla_p z \quad \text{Eq. 7-7}$$

$$\begin{cases} u_g = -\frac{1}{f\rho}\frac{\partial p}{\partial y} \\ v_g = \frac{1}{f\rho}\frac{\partial p}{\partial x} \end{cases}$$

Compare:
- **Thermal wind** is determined by the **horizontal temperature gradient**,
- **geostrophic wind** through the **horizontal pressure gradient**.

Thermal wind – examples

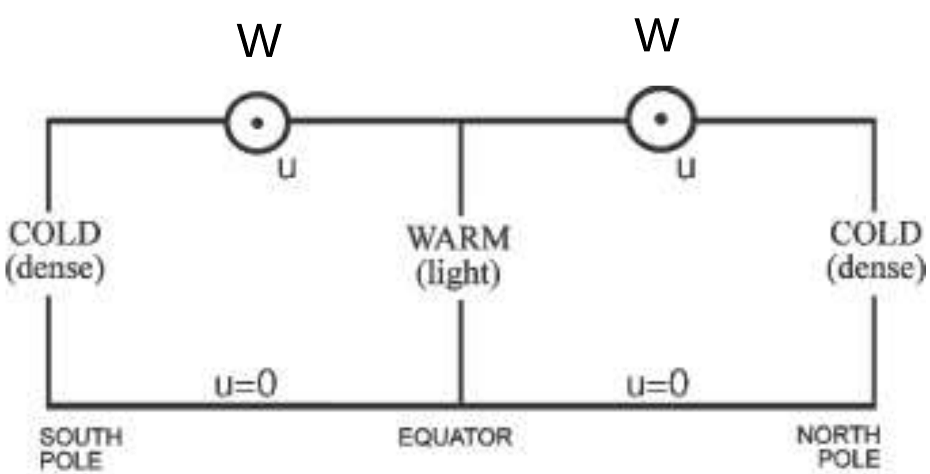
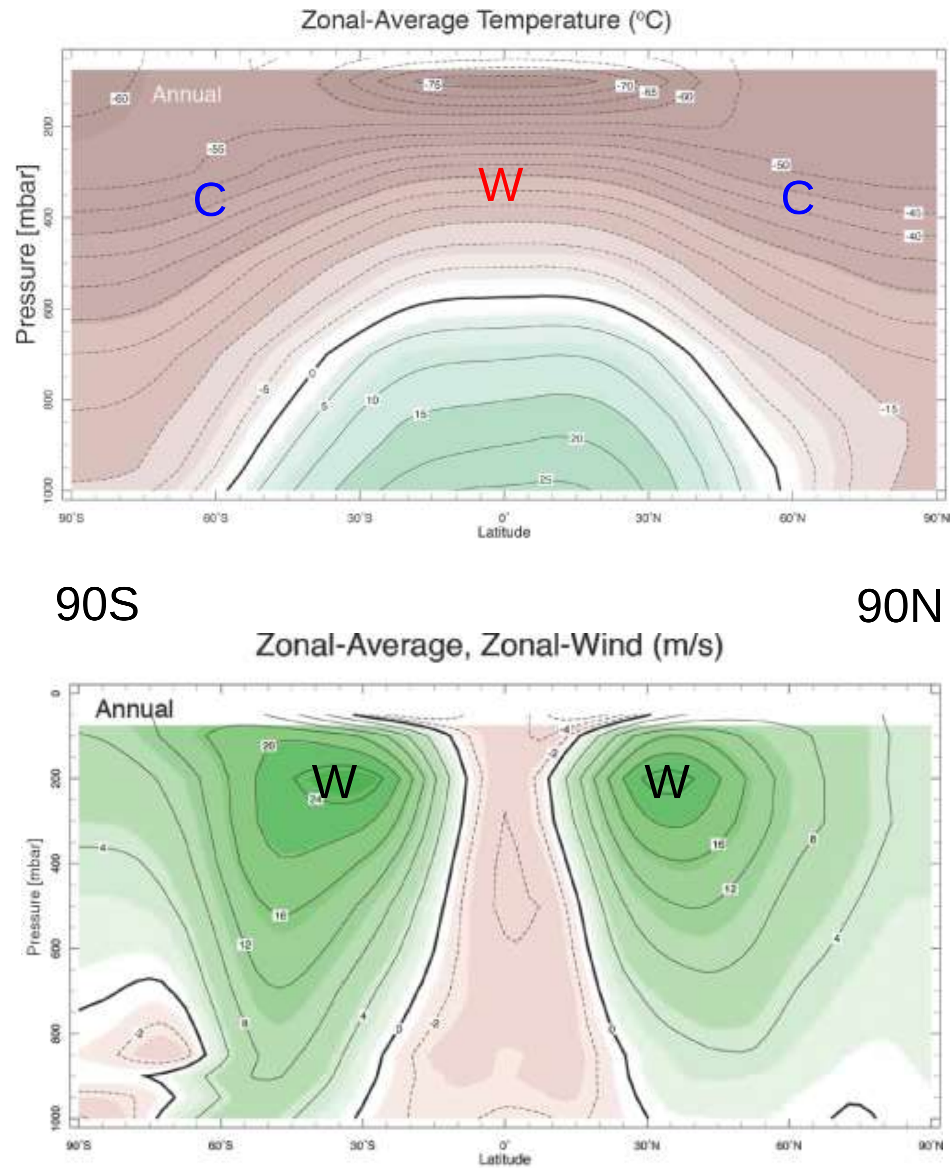


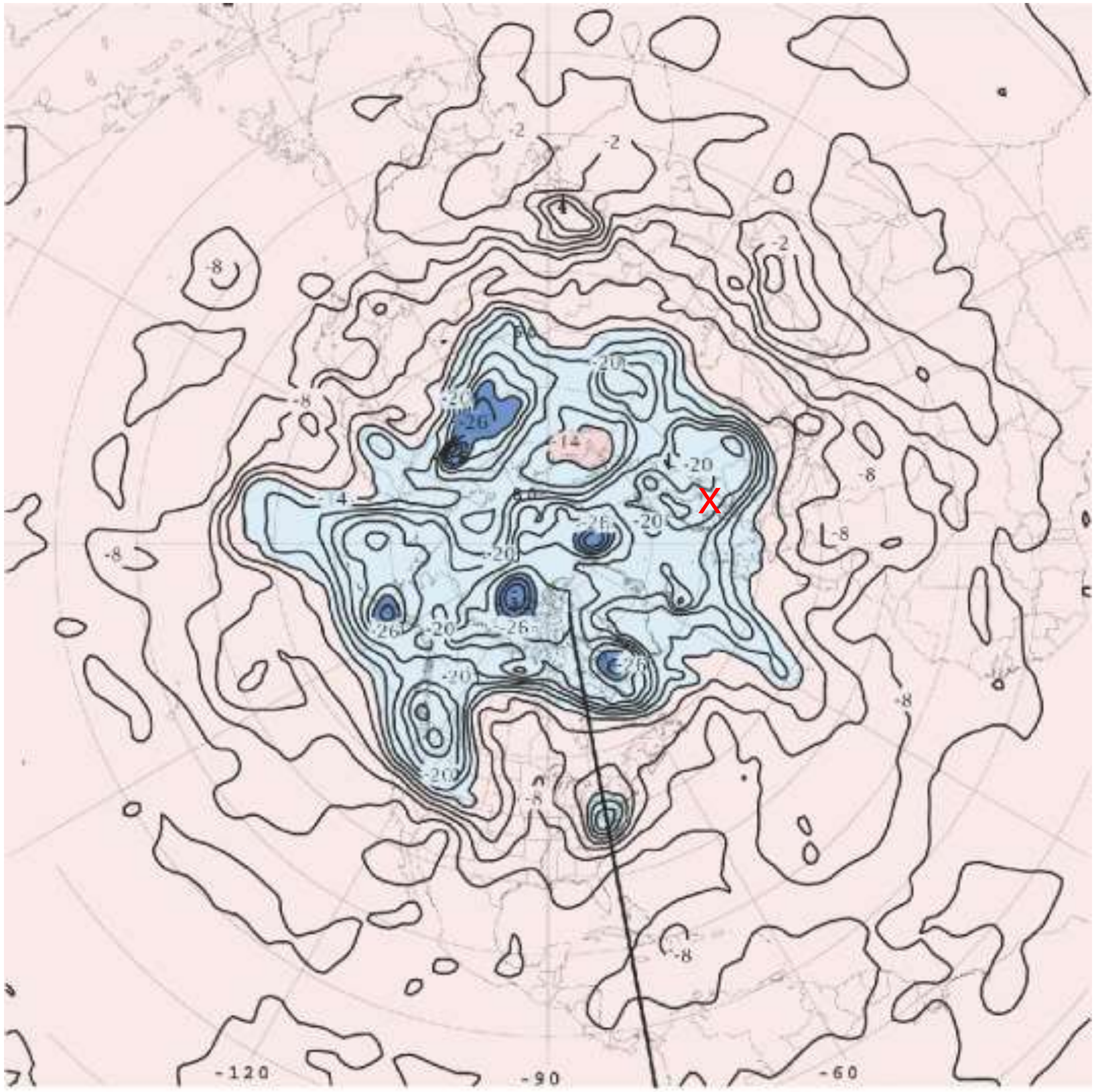
Figure 7.19: A schematic of westerly winds observed in both hemispheres in thermal wind balance with the equator-to-pole temperature gradient. (See Eq. 7-24 and the observations shown in Figs. 5.7 and 5.20.)



Marshall and Plumb (2008)

Temperature (°C) @500hPa, 12 GMT June 21, 2003

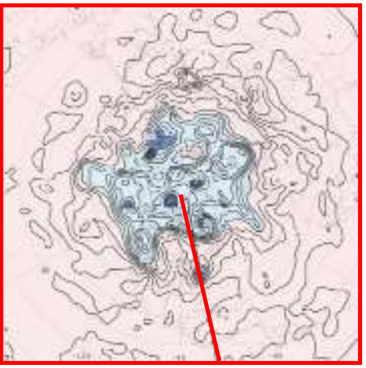
cold air
warm air

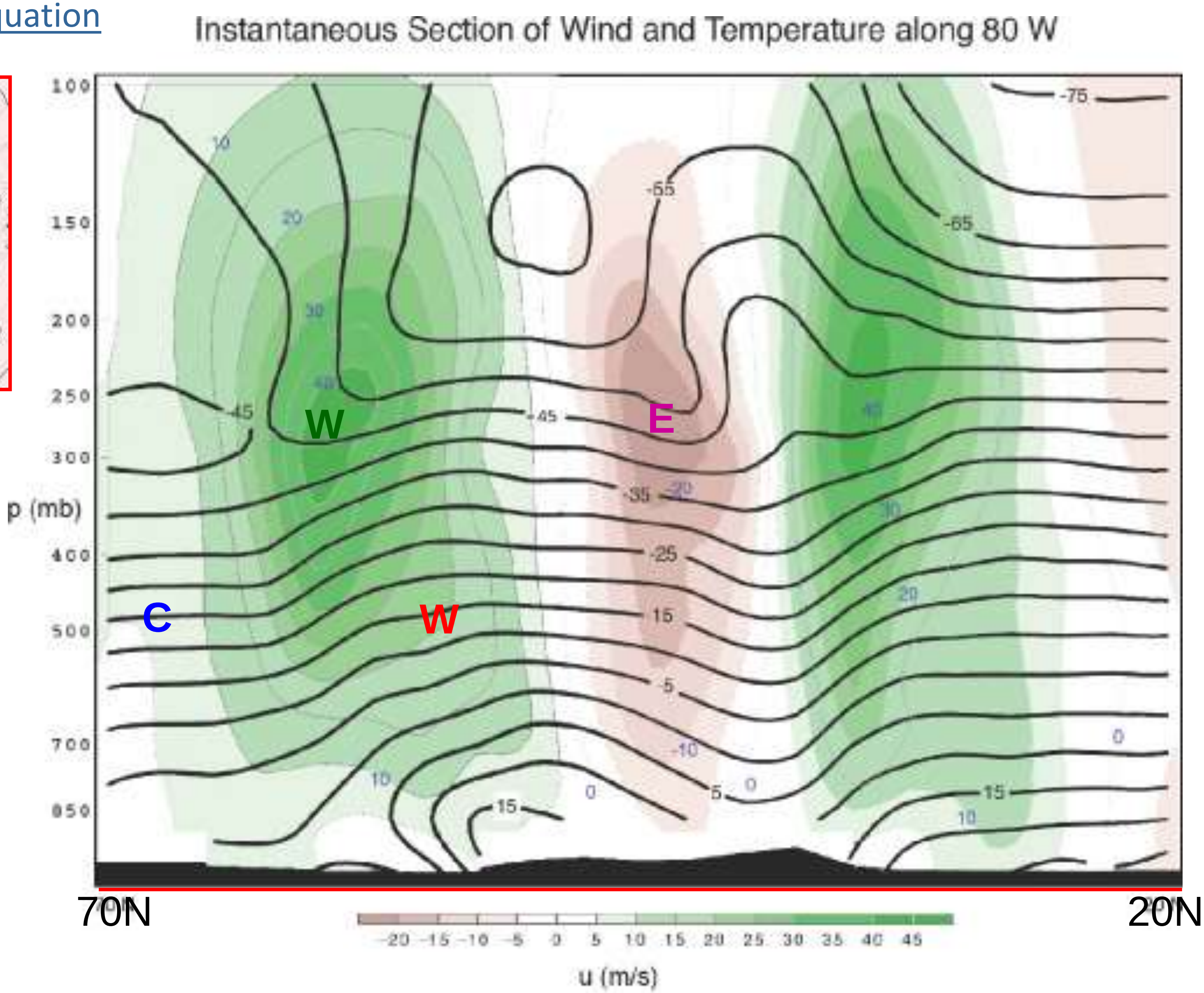


x Oslo

Figure 7.20: The temperature, T, on the 500-mbar surface at 12 GMT on June 21, 2003, the same time as Fig. 7.4. The contour interval is 2°C. The thick black line marks the position of the meridional section shown in Fig. 7.21 at 80°W extending from 20° N to 70° N. A region of pronounced temperature contrast separates warm air (pink) from cold air (blue). The coldest temperatures over the pole get as low as –32°C.

Marshall and Plumb (2008)





Westerly wind
Easterly wind

Warm
Cold

Figure 7.21: A cross section of zonal wind, u (color-scale, green indicating away from us and brown toward us, and thin contours every 5 m s^{-1}), and temperature, T (thick contours every 5°C), through the atmosphere at 80° W, extending from 20° N to 70° N, on June 21, 2003, at 12 GMT, as marked on Figs. 7.20 and 7.4. Note that $\partial u/\partial p < 0$ in regions where $\partial T/\partial y < 0$ and vice versa.

Marshall and Plumb (2008)

Stratospheric polar vortex

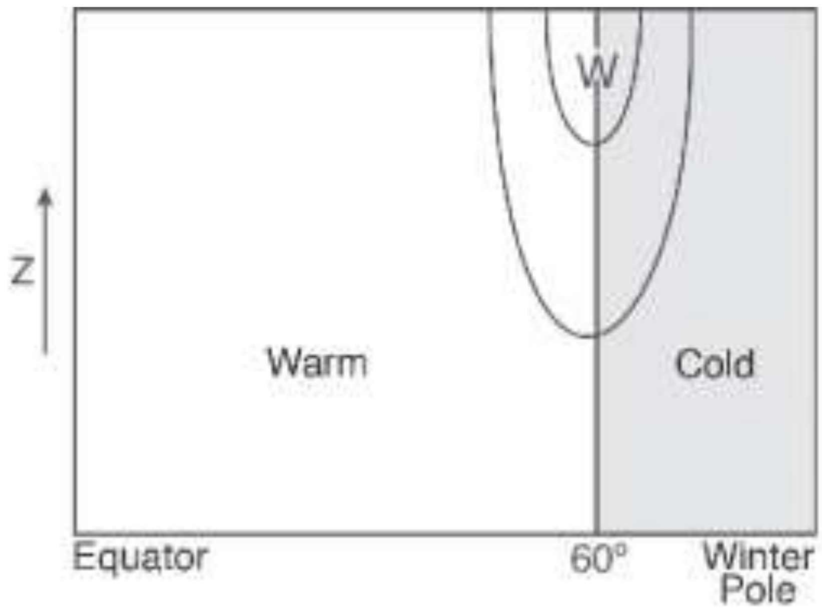


Figure 7.29: A schematic of the winter polar stratosphere dominated by the "polar vortex," a strong westerly circulation at ~60° around the cold pole.

Marshall and Plumb (2008)

Sub-geostrophic flow: The Ekman layer

- **Large scale flow** in **free atmosphere** and **ocean** is close to **geostrophic** and **thermal wind balance**.
- Boundary layers have large departures from geostrophy due to friction forces.

Ekman layer:

- Friction acceleration $\mathcal{F}$ becomes important,
- roughness of surface generates turbulence in the first ~1 km of the atmosphere,
- wind generates turbulence at the ocean surface down to first 100 meters.

Sub-geostrophic flow:

R_0 small, $\mathcal{F}$ exists, then horizontal component of momentum (geostrophic) balance (Eq. 7-2) can be written as:

$$f\hat{\boldsymbol{z}} \times \boldsymbol{u} + \frac{1}{\rho}\nabla p = \mathcal{F} \qquad \text{Eq. 7-25}$$

Surface (friction) wind

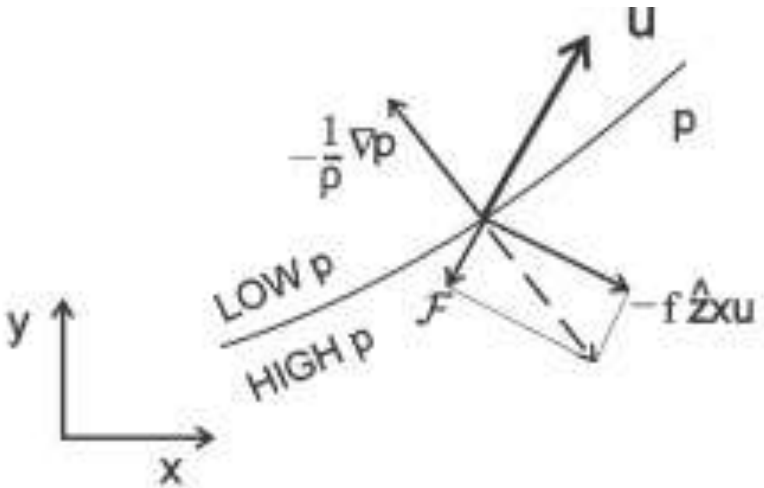


Figure 7.22: The balance of forces in Eq. 7-25: the dotted line is the vector sum $\mathcal{F} - f\hat{\mathbf{z}} \times \boldsymbol{u}$ and is balanced by $-1/\rho\, \nabla p$.

$$f\hat{\mathbf{z}} \times \boldsymbol{u} + \frac{1}{\rho}\nabla p = \mathcal{F}$$ Eq. 7-25

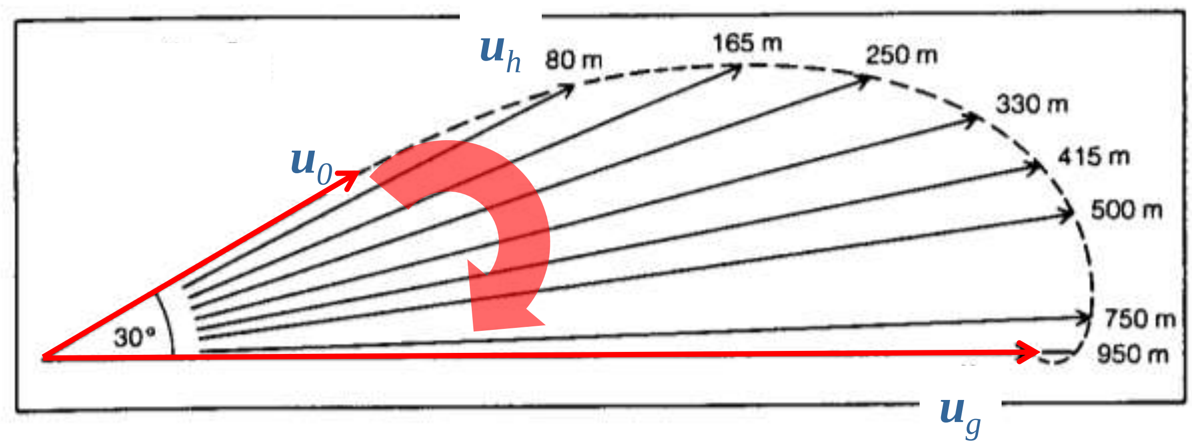
Marshall and Plumb (2008)

1. Start with ***u***
2. Coriolis force per unit mass must be to the right of ***u***
3. Frictional force per unit mass acts as a drag and must be opposite to ***u***.
4. Sum of the two forces (dashed arrow) must be balanced by the pressure gradient force per unit mass
5. Pressure gradient is not normal to the wind vector or the wind is no longer directed along the isobars, however the low pressure system is still on the left side but with a deflection.

=> The flow is sub-geostrophic (less than geostrophic); ageostrophic component directed from high to low.

The Ekman spiral – theory

Ekman spiral: simplified theoretical calculation



Roedel, 1987

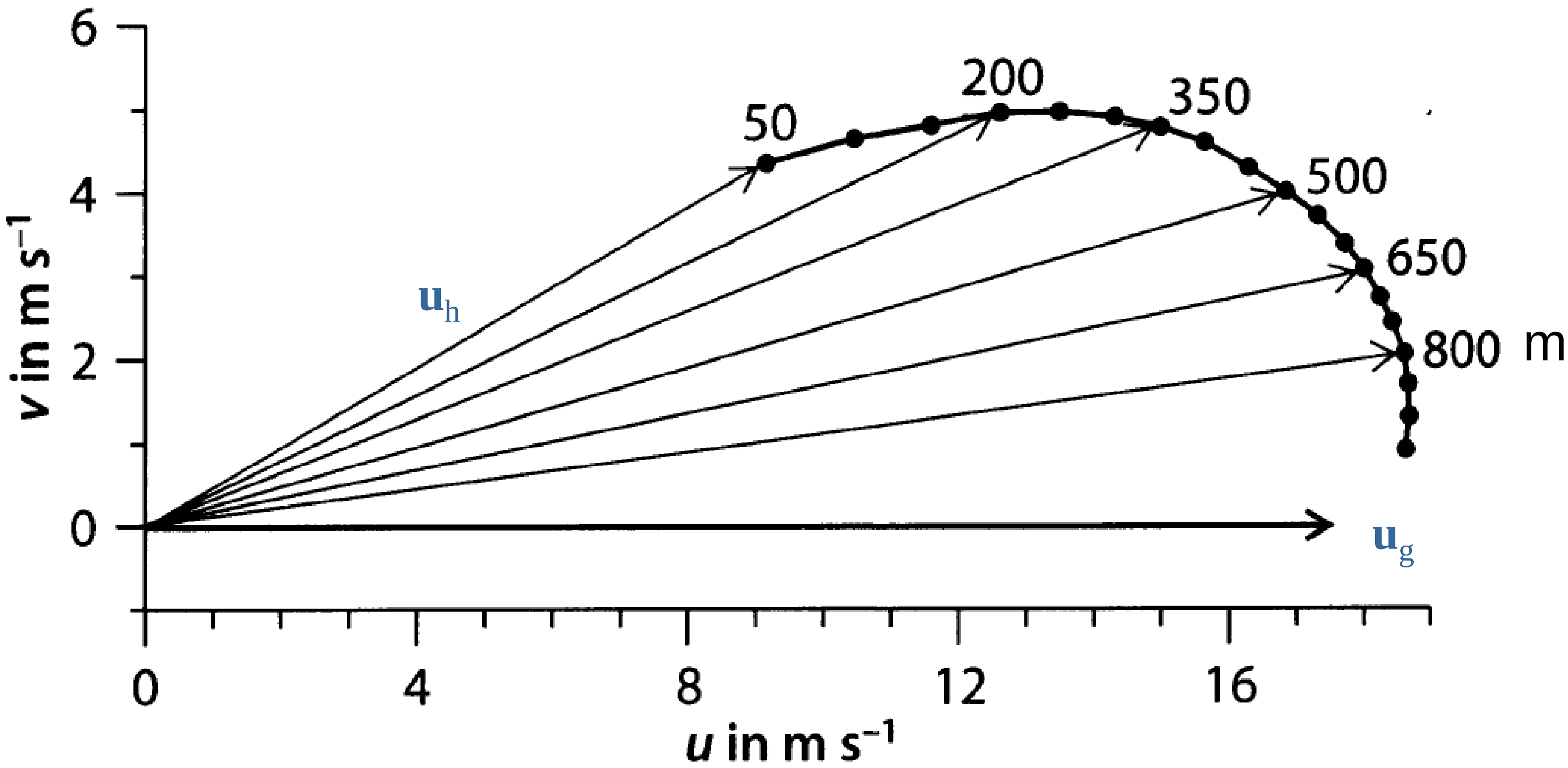
$$f\hat{z} \times \boldsymbol{u} + \frac{1}{\rho}\nabla p = \mathcal{F}$$ Eq. 7-25

$\mathcal{F} = \mu\, \partial^2 \boldsymbol{u} / \partial z^2$

μ: constant eddy viscosity

The Ekman spiral was first calculated for the oceanic friction layer by the Swedish oceanographer in 1905. In 1906, a possible application in Meteorology was developed.

Wind measurement in the boundary layer



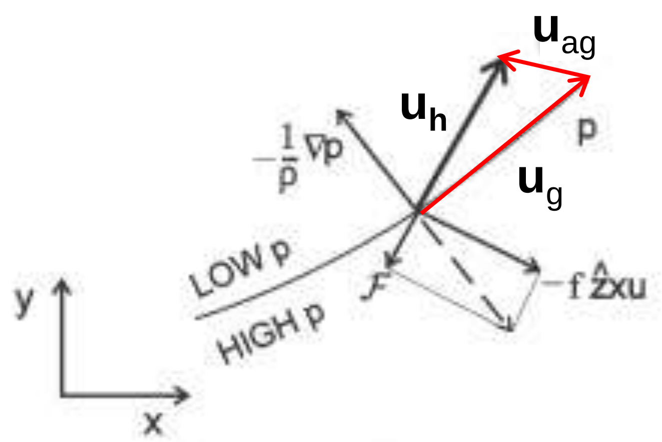
The “Leipzig wind profile” Kraus, 2004

The ageostrophic flow $\mathbf{u}_{ag}$

The *ageostrophic* flow is the difference between the geostrophic flow ($\mathbf{u}_g$) and the horizontal flow ($\mathbf{u}_h$):

$\mathbf{u}_h = \mathbf{u}_g + \mathbf{u}_{ag}$ Eq. 7-26

$f\hat{\mathbf{z}} \times \boldsymbol{u}_{ag} = \mathcal{F}$ Eq. 7-27



The ageostrophic component is always directed to the right of $\mathcal{F}$ in the NH.

Surface (friction) wind

Low pressure (NH):
- anticlockwise flow
- often called "cyclone"

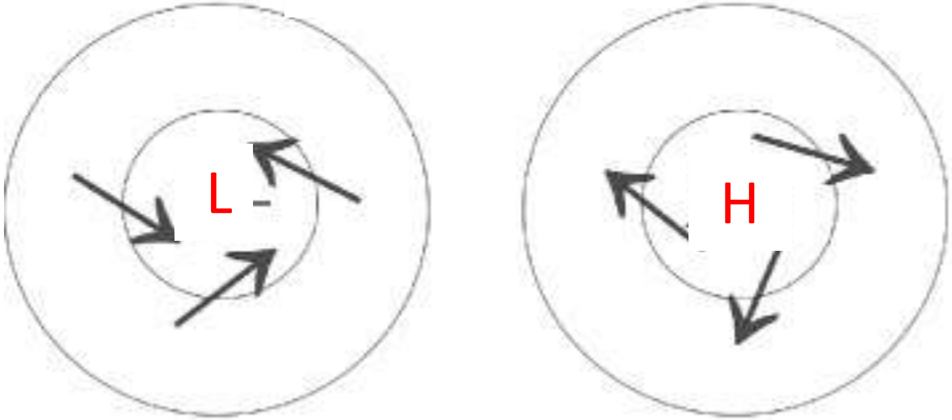


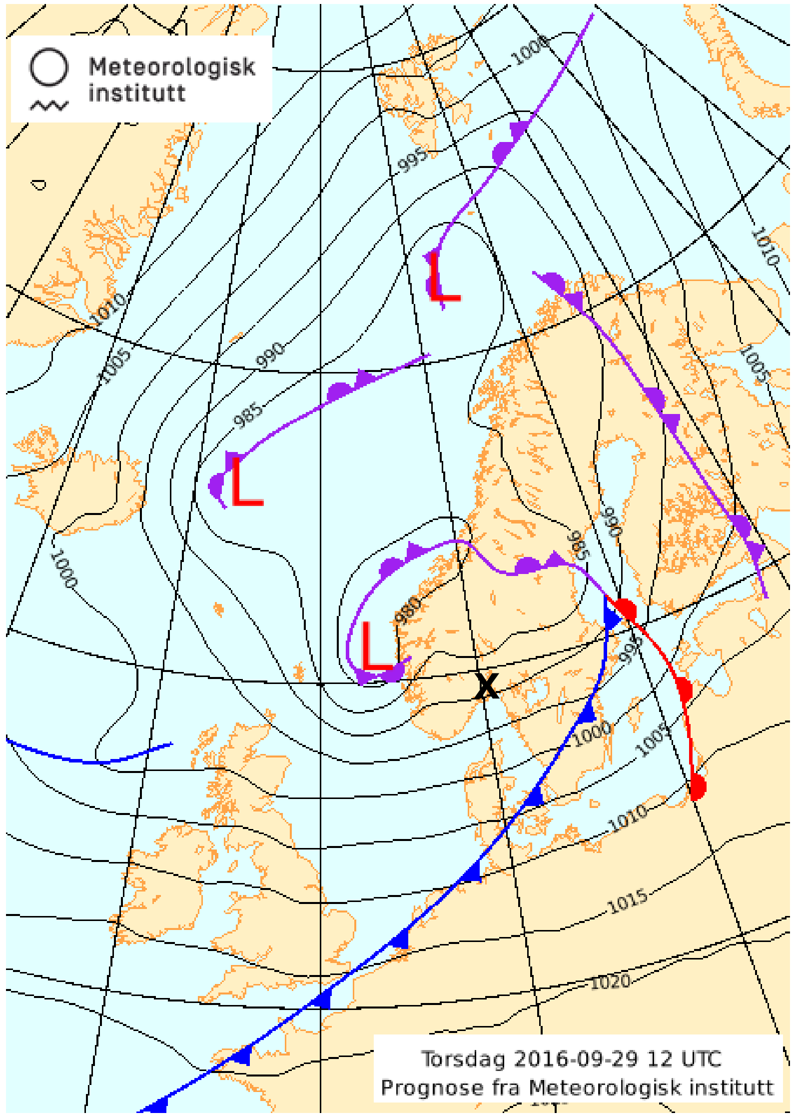
Figure 7.24: Flow spiraling in to a low-pressure region (left) and out of a high-pressure region (right) in a bottom Ekman layer. In both cases the ageostrophic flow is directed from high pressure to low pressure, or down the pressure gradient.

High pressure (NH):
- clockwise flow
- often called "anticyclone"

Marshall and Plumb (2008)

Note: At the surface the wind is blowing from the high to the low pressure system.

Today‘s weather chart



x Oslo

Varmfront

Kaldfront

Okklusjon

What is the wind direction and approximately wind strength in Oslo?

a) Weak northeasterly.
b) Fresh southwesterly.
c) Storm from the west.
d) Storm from the east.

www.yr.no

Beaufort wind scale

«Sir Francis Beaufort var en irsk hydrograf og offiser i den britiske marinen. I 1806 satte han navn på vindens styrke. Denne skalaen brukes i dag i all vanlig værvarsling.»

http://om.yr.no/forklaring/symbol/vind/

Vindens virkning på sjøen

Navn	Symbol	m/s	knop	Kjennetegn
Stille	○	0,0-0,2	0-1	Sjøen er speilblank (havblikk).
Flau vind		0,3-1,5	1-3	Vindretning sees av røykens drift.
Svak vind		1,6-3,3	4-6	Små korte, men tydelige bølger med glatte kammer som ikke brekker.
Lett bris		3,4-5,4	7-10	Småbølgene begynner å toppe seg, det dannes skum, som ser ut som glass. en og annen skumskavl kan forekomme.
Laber bris		5,5-7,9	11-16	Bølgene blir lengre, endel skumskavler.
Frisk bris		8,0-10,7	17-21	Middelstore bølger som har mer utpreget langstrakt form og med mange skumskavler. Sjøsprøyt fra toppene kan forekomme.
Liten kuling		10,8-13,8	22-27	Store bølger begynner å danne seg. Skumskavlene er større overalt. Gjerne noe sjøsprøyt.
Stiv kuling		13,9-17,1	28-33	Sjøen hoper seg opp og hvitt skum fra bølgetopper som brekker, begynner å blåse i strimer i vindretningen.
Sterk kuling		17,2-20,7	34-40	Middels høye bølger av større lengde. Bølgekammene er ved å brytes opp til sjørokk, som driver i tydelige markerte strimer med vinden.
Liten storm		20,8-24,4	41-47	Høye bølger. Tette skumstrimer driver i vindretningen. Sjøen begynner å rulle. Sjørokket kan minske synsvidden.
Full storm		24,5-28,4	48-55	Meget høye bølger med lange overhengende kammer. skummet, som dannes i store flak, driver med vinden i tette hvite strimer så sjøen får et hvitaktig utseende. Rullingen blir tung og støtende. Synsvidden nedsettes.
Sterk storm		28,5-32,6	56-63	Ualminnelig høye bølger (små og middelstore skip kan for en tid forsvinne i bølgedalene). Sjøen er fullstendig dekket av lange, hvite skumflak som ligger i vindens retning. Overalt blåser bølgekammene til frådelignende skum. Sjørokket nedsetter synsvidden.
Orkan		32,6-	64-	Luften er fylt av skum og sjørokk som nedsetter synsvidden betydelig. Sjøen er fullstendig hvit av drivende skum.

Vertical motion induced by Ekman layers

- Ageostrophic flow is horizontally divergent.
- Convergence/divergence drives vertical motions.
- In pressure coordinates, if *f* is const., the continuity equation (Eq. 6-12) becomes:

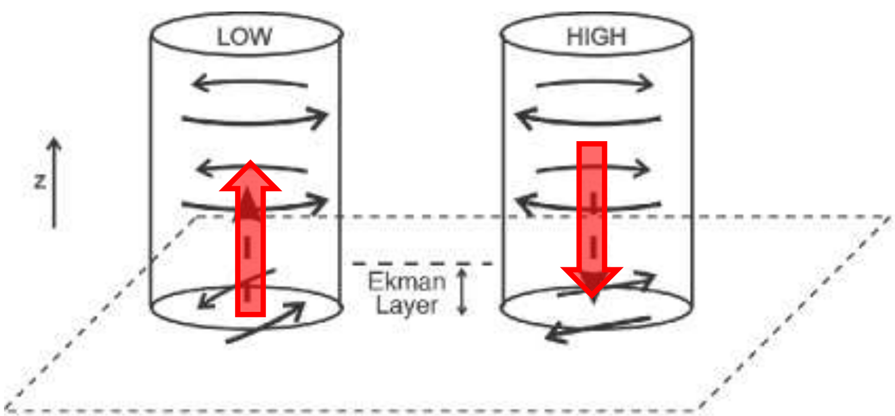


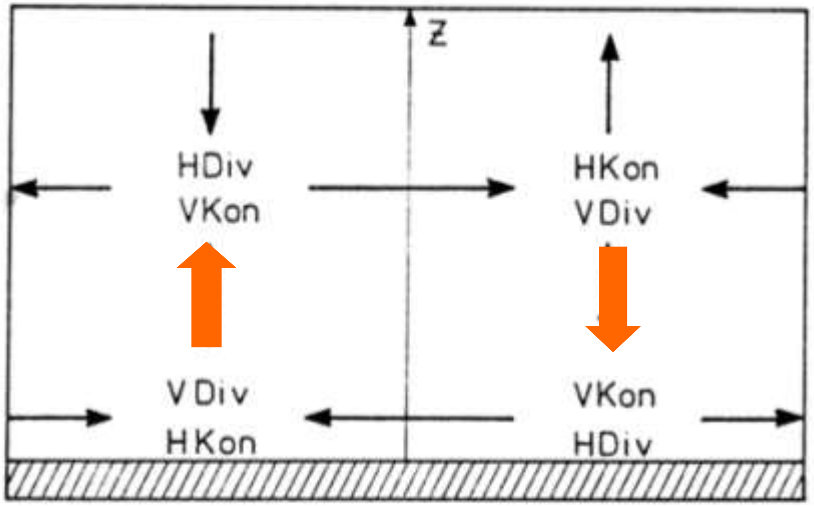
Figure 7.26: Schematic diagram showing the direction of the frictionally induced ageostrophic flow in the Ekman layer induced by low pressure and high pressure systems. There is flow into the low, inducing rising motion (the dotted arrow), and flow out of a high, inducing sinking motion.

Marshall and Plumb (2008)

$$\nabla_p \cdot \boldsymbol{u}_{ag} + \frac{\partial w}{\partial p} = 0$$

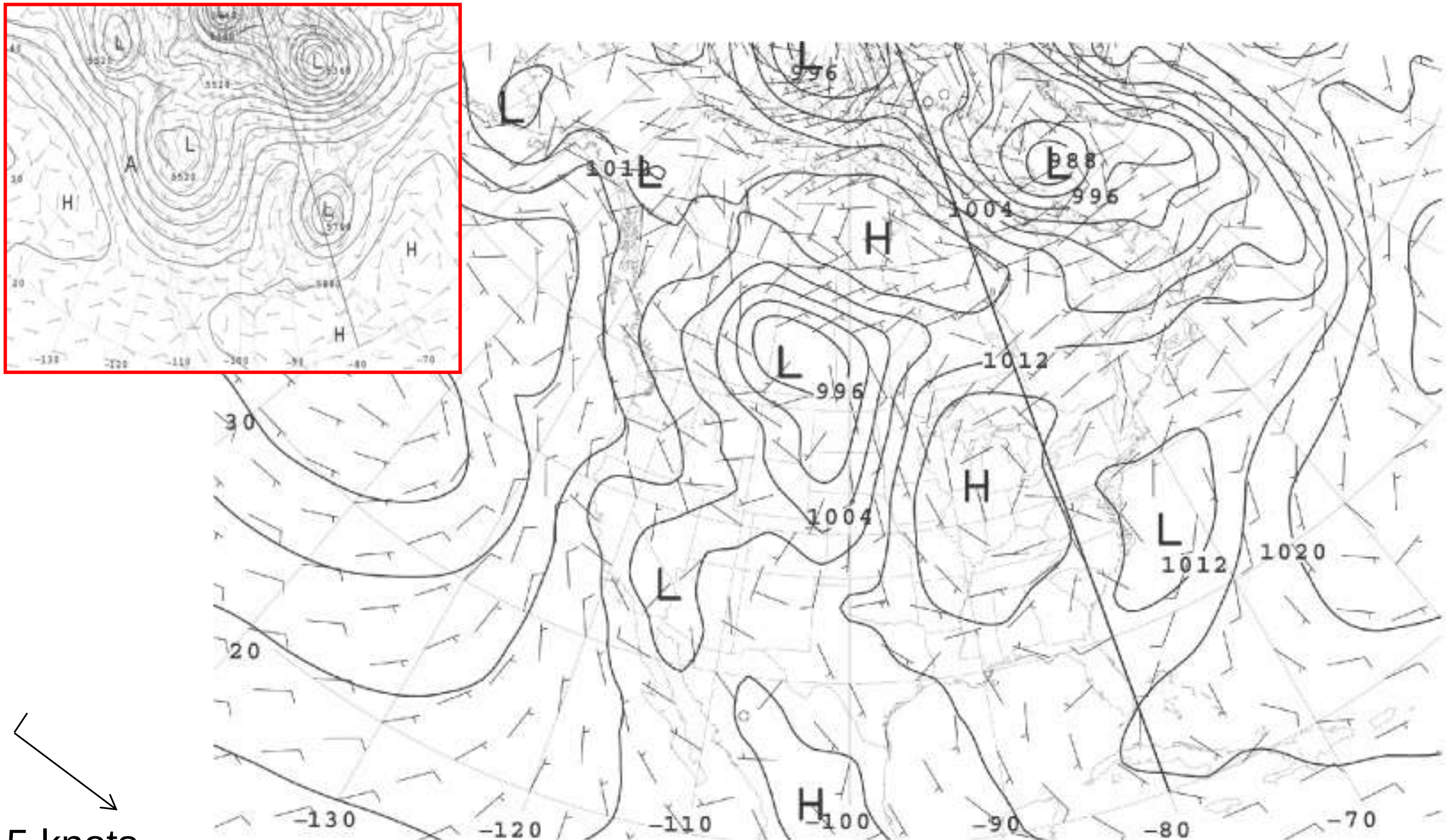
- Low: convergent flow
- High: divergent flow
- Continuity equation requires vertical motions:
  - "Ekman pumping" > ascent
  - "Ekman suction" > descent

Characteristics of horizontal wind



Combination of horizontal and vertical vergences

Surface pressure (hPa) and wind (kn), 12 GMT June 21, 2003, USA

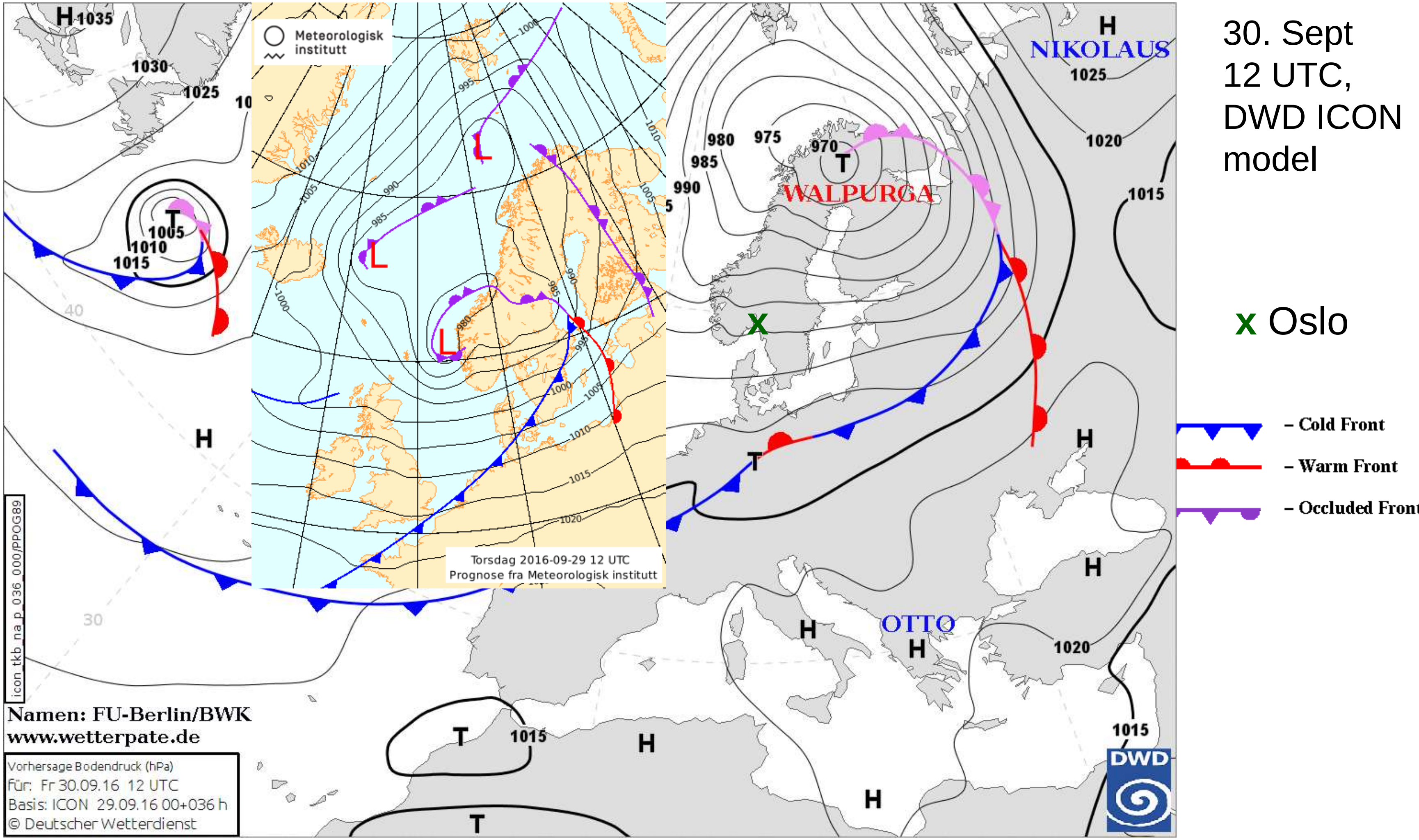


5 knots

10 knots

Marshall and Plumb (2008)

Tomorrow’s surface pressure (hPa) map



30. Sept
12 UTC,
DWD ICON
model

x Oslo

www.met.fu-berlin.de

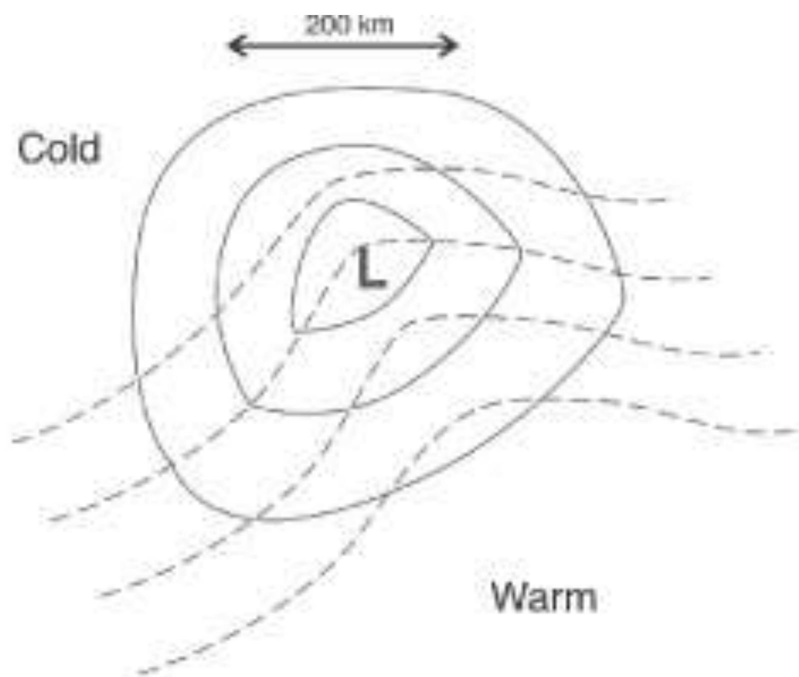


Figure 7.30: A schematic of surface pressure contours (solid) and mean 1000 mbar – – – 500 mbar temperature contours (dashed), in the vicinity of a typical northern hemisphere depression (storm).

See also chapter 8.

Marshall and Plumb (2008)

Summary of chapters 6 and 7

TABLE 7.1. Summary of key equations. Note that (x, y, p) is not a right-handed coordinate system. So although $\hat{\mathbf{z}}$ is a unit vector pointing toward increasing z, and therefore upward, $\hat{\mathbf{z}}_p$ is a unit vector pointing toward decreasing p, and therefore also upward.

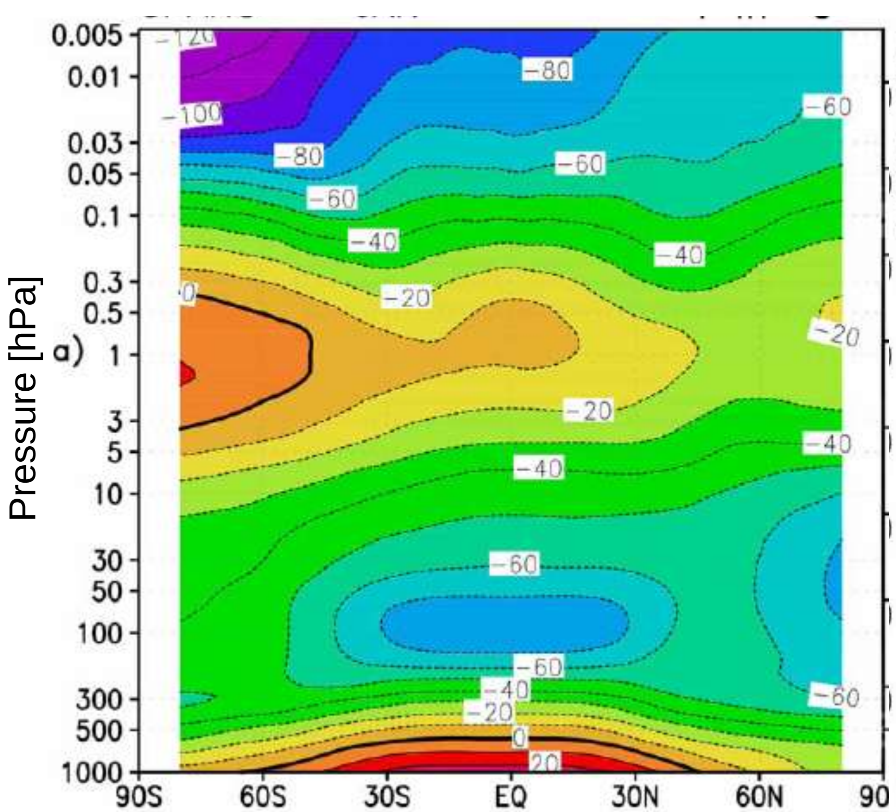
(x, y, z) coordinates	(x, y, z) coordinates	(x, y, p) coordinates
$\nabla \equiv \left(\frac{\partial}{\partial x}, \frac{\partial}{\partial y}, \frac{\partial}{\partial z}\right)$ general	$\nabla \equiv \left(\frac{\partial}{\partial x}, \frac{\partial}{\partial y}, \frac{\partial}{\partial z}\right)$ (incompressible—OCEAN)	$\nabla_p \equiv \left(\frac{\partial}{\partial x}, \frac{\partial}{\partial y}, \frac{\partial}{\partial p}\right)$ (comp. perfect gas—ATMOS)
Continuity $\frac{\partial \rho}{\partial t} + \nabla \cdot (\rho u) = 0$	$\nabla \cdot u = 0$	$\nabla_p \cdot u = 0$
Hydrostatic balance $\frac{\partial p}{\partial z} = -g\rho$	$\frac{\partial p}{\partial z} = -g\rho$	$\frac{\partial z}{\partial p} = -\frac{1}{g\rho}$
Geostrophic balance $fu = \frac{1}{\rho}\hat{z} \times \nabla p$	$fu = \frac{1}{\rho_{ref}}\hat{z} \times \nabla p$	$fu = g\hat{z}_p \times \nabla_p z$
Thermal wind balance	$f\frac{\partial \boldsymbol{u}}{\partial z} = ag\, \hat{\boldsymbol{z}} \times \nabla T$	$f\frac{\partial u}{\partial p} = -\frac{R}{p}\hat{z}_p \times \nabla T$

Take home message

- Balanced flows for horizontal fluids (atmosphere and ocean).
- Balanced horizontal winds:
  - **Geostrophic wind balance** good approximation for the observed wind in the free troposphere.
  - **Gradient wind balance** and **cyclostrophic wind balance** occur with higher Rossby number.
  - **Surface wind** (subgeostrophic flow) **balance** occurs within the Ekman layer.
- **Thermal wind** is good **approximation** for the geostrophic wind change with height *z*.

Which winds play a role in the below shown climatology?

Zonal mean **temperature** (° C)



Zonal mean **zonal wind** (m/s)